THE EFFECTS OF METAMORPHISM, TECTONICS AND HOST ROCKS ON THE LOCATION OF SULPHIDE ORES IN THE KONGSBERG SERIES, SOUTH NORWAY.

by

TIMOTHY ERIC DIBB

Queen Mary College

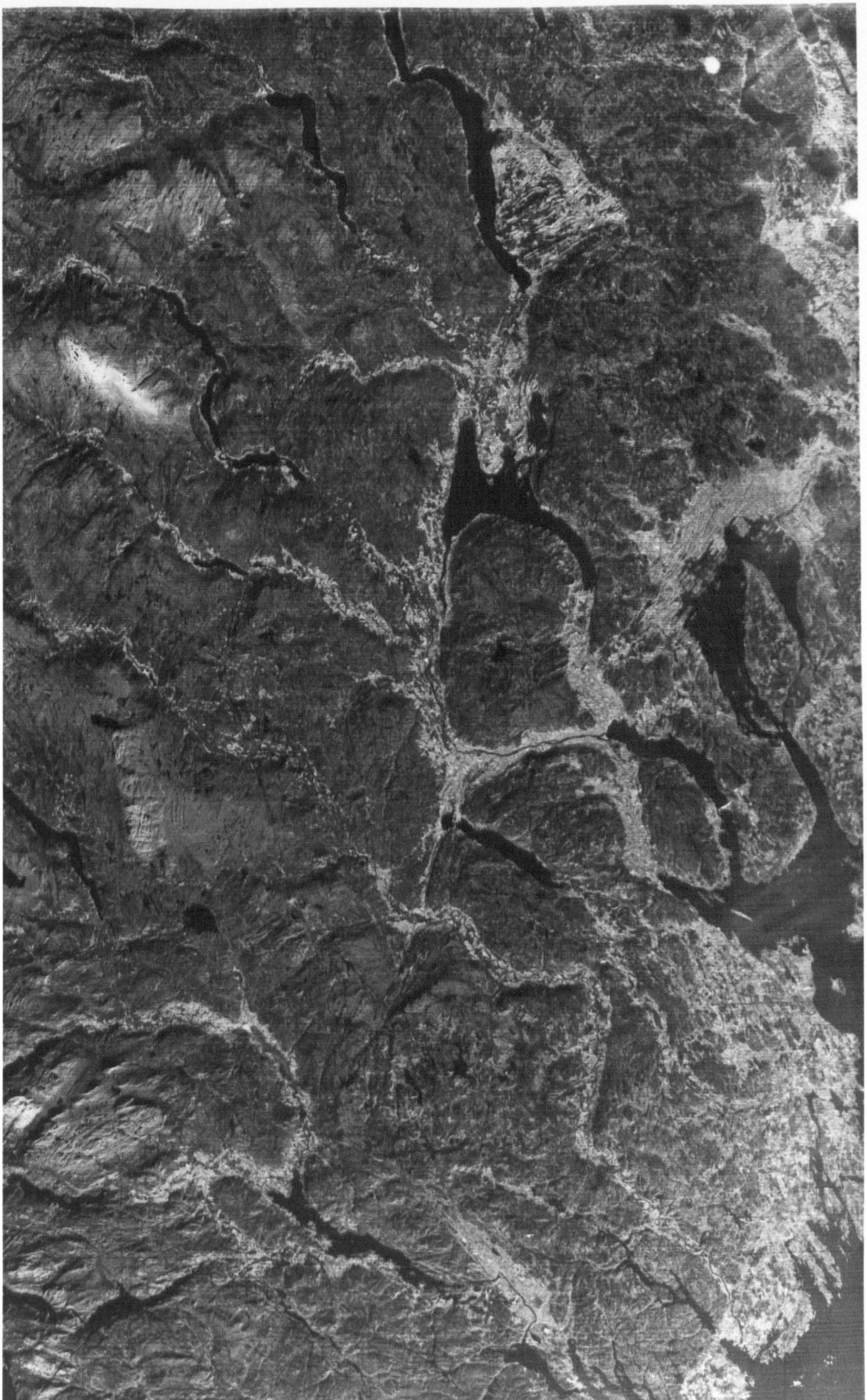
University of London

Thesis presented for the degree of Doctor of Philosophy

in the University of London

CONTAINS

PULLOUTS



ACKNOWLEDGEMENTS

I wish to thank Dr. I. C. Starmer for his supervision and help. The research was funded by a NERC/CASE studentship in association with Norsk Hydro A/S: their help in making data available and financing the field seasons is gratefully acknowledged. I would also like to thank Professor F. M. Vokes for inviting me to study for six weeks in Trondheim and for his critical reading of several chapters. Professors G. Kullerud and R. L. Stanton are thanked for their constructive discussions. Thanks are also due to Dr. J. Pearce for his advice and permission to use unpublished diagrams.

I would like to thank Professor M. G. Audley-Charles for allowing the use of facilities at Queen Mary College, and the technical staff of the Department of Geological Sciences for their help; especially Messrs. D. Lawrence, K. Schrapel, A. McLachlan, S. Adams and C. Cook. I am also very grateful to Mrs. J. Schrapel for typing the manuscript. Finally, thanks to Jane for all her support, both moral and financial, during the writing of the thesis.

ABSTRACT

A study has been made of two deposits of Fe-Cu-Zn sulphide ores, from Grøsli and Eiker, in the Proterozoic Kongsberg Series of South Norway. The ores are located at the junction of acid-intermediate supracrustals and amphibolitized gabbros. The supracrustals underwent Upper Amphibolite facies metamorphism of Svecofennian age (1600 to 1500 m.y.B.P.) and were then intruded by the gabbros which underwent subsequent Sveconorwegian (1200 to 1000 m.y.B.P.) metamorphism at Mid-Amphibolite facies grade.

The ore bodies were originally part of the supracrustal sequence, which was deposited as a volcano-sedimentary succession (with some hydrothermally altered equivalents), similar to the Kuroko-type deposits of Japan. The ores have thus undergone both phases of high grade regional metamorphism. At Grøsli, the ores were partially incorporated in the gabbros prior to the later (Sveconorwegian) metamorphism, during which extensive shearing occurred throughout the ores at Eiker.

The cores of the gabbroic intrusions retained original igneous mineralogies and textures, with progressive amphibolitization towards the peripheries. A subsequent Epidote-Amphibolite and later Greenschist facies grade overprint produced varying retrograde assemblages in the silicate rocks. Under Greenschist facies conditions, rejuvenation of the Sveconorwegian shear (at Eiker) caused brittle faulting, while fluid activity (at both Grøsli and Eiker) caused the formation of chlorite-actinolite-carbonate assemblages around some of the ore bodies. Re-mobilisation of the ore material itself was minimal.

Studies of geothermometry and geobarometry indicate that the ore

deposits were re-equilibrated during the Epidote-Amphibolite facies overprinting.

CONTENTS

Page No.

CHAPTER THREE:

PETROLOGY OF THE SILICATE ROCKS AT GRØSLI

Page No.

Page No.

CHAPTER SIX:

STRUCTURE, LITHOLOGIES AND AGE RELATIONS AT EIKER

CHAPTER SEVEN:

PETROLOGY OF THE SILICATE ROCKS AT EIKER

CHAPTER EIGHT:

THE ORE DEPOSIT AT EIKER AND ITS METAMORPHISM

Page No.

CHAPTER NINE:

CHEMISTRY OF THE SILICATE ROCKS AT EIKER

CHAPTER TEN:

GEOTHERMOMETRY AND GEOBAROMETRY OF THE ORE DEPOSITS AT GRØSLI AND EIKER

CHAPTER ELEVEN:

THE ORIGINS OF THE ORE DEPOSITS

Page No.

CHAPTER ONE

GENERAL INTRODUCTION AND REGIONAL GEOLOGY

1.1 SCOPE OF THE PROJECT

Detailed investigations have been carried out on two Fe-Cu-Zn sulphide ore deposits within the Kongsberg Series of the Fennoscandian shield. The relationships between the ore deposits and the surrounding Proterozoic rocks have been examined.

To the west of the Phanerozoic Oslo Graben, the Precambrian of South Norway comprises three complexes; the large central block of the Telemark Series (Fig. 1.1) is bordered to the east by the Kongsberg Series and to the south-east by the Bamble Series. The Kongsberg and Bamble Series have been previously grouped together as the 'Kongsberg-Bamble-formasjon' but whilst their geological history shows some similarities, they comprise somewhat different lithologies (Starmer 1977).

The ore deposits studied both lie within the Kongsberg Series block (Fig. 1.2). The Grøsli mine lies in the west of the Series about five kilometres from the boundary with the Telemark Series. The Eiker mine is situated some twenty kilometres south-east of Grøsli, three to four kilometres from the eastern boundary with the Oslo Graben.

Detailed geological maps have been prepared for the areas around the Grøsli and Eiker mines. Data from twelve boreholes at Grøsli enabled a detailed interpretation of the sub-surface geology. Extensive sampling was also carried out: both surface and borehole samples were sectioned and polished for microscope studies.

Ores and surrounding silicate rocks were analysed for major and

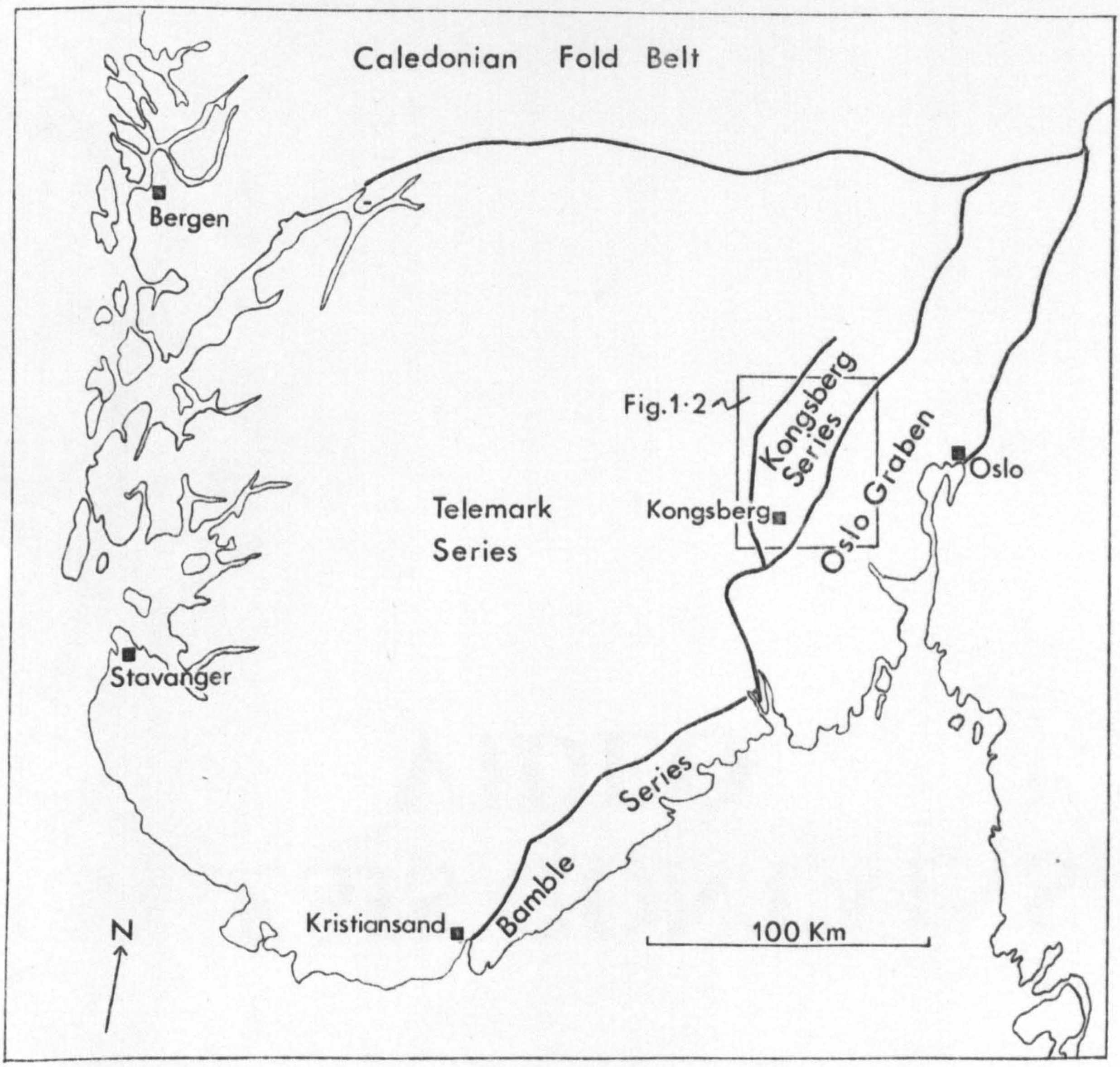


FIG.1·1 Tectonic divisions of South Norway

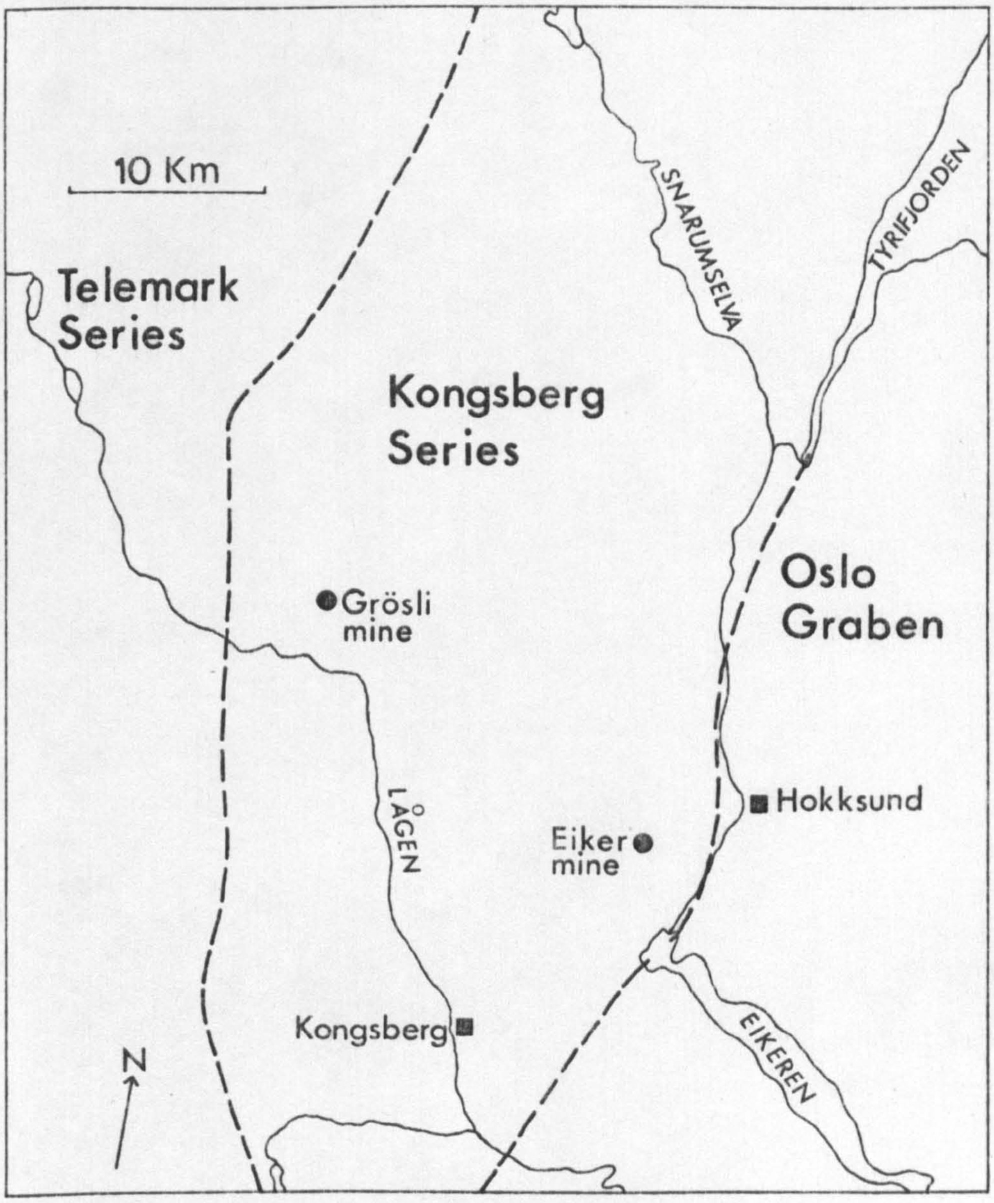


FIG.1·2 Location of mine areas

trace elements. The silicate rocks were quantitatively analysed for thirteen traces (Cu, Zn, Cr, Ni, Co, Cd, V, Ce, Ba, Zr, Y, Sr and Rb). The ores were analysed for eleven elements (Fe, Cu, Zn, Cr, Ni, Co, Cd, Pb, V, Mn and Ti). Analyses were performed by X-ray fluorescence spectrometry and plasma-arc spectrometry using international standards. Analyses of samples from Grøsli and Eiker using standard wet chemical methods were used to check results.

Electron microprobe analyses of the ores were also carried out for geothermometric-geobarometric studies (Chapter Ten). In addition electron microprobe analyses of silicates (olivines, pyroxenes, amphiboles, biotites, chlorites and garnets) were carried out to determine their exact composition.

Chapters Two to Nine include the presentation of the data from each mine in turn. Chapter Ten discusses the temperatures and pressures of the metamorphism of the ore deposits. Chapter Eleven reviews possible origins of the ore deposits in relation to theories of ore deposition. Chapter Twelve compares the environments of Proterozoic and Phanerozoic ore formation and analyses the spatial relationships between the two ore bodies studied.

1.2 GENERAL REGIONAL GEOLOGY

Early studies on the rocks in the Kongsberg Series were carried out by C. Bugge (1917) and A. Bugge (1937). These were later summarized by J. Bugge (1943). Recent work has been carried out in the south-western and western parts of the Series and across the junction with the Telemark Series (Starmer, 1977, 1979).

The Kongsberg Series comprises a group of supracrustal gneisses with major intrusive granitic, dioritic and basic lithologies of different ages. The oldest rocks have been affected by Svecofennian, Upper Amphibolite grade metamorphism. The Series was subsequently cataclased (especially in the western part) and then suffered Sveconorwegian Mid-Amphibolite facies metamorphism.

Descriptions of the lithologies have been approached in various ways by different workers. A. Bugge (1937) described the supracrustal lithologies and the younger intrusives ('Yngre eruptivbergarter'), comprising 'Vinor' dolerites, olivine hyperites and gabbros, together with granites and pegmatites.

J. Bugge (1943) introduced a concept of a 'metamorphic climax' or 'migmatization period' to divide the 'Young' and 'Old' rock groups. The 'Old Complex' consisted of metasediments and metavolcanics (quartzites, mica schists, sillimanite gneisses, limestones, banded gneisses and 'effusives') and early intrusives ('plutonics') together with later gabbroic bodies ('hyperites'). Metamorphism and metasomatism then migmatized all these rocks to produce the 'Young Complex'. Bugge used examples largely from the Bamble region on the Skagerrak coast, but stated that "in the Kongsberg formasjon we find all connecting links between the metamorphic and migmatic rocks". Rocks of the 'Old Complex' were progressively granitized and boundaries between acid and basic gneiss bands became obliterated, "with acid ptygmas developed in amphibolites". Granites were also produced by ascending 'ichors' rich in silica and alkalies.

At the western margin of the Kongsberg Series, A. Bugge (1937) noted that; "A great friction breccia, 100 to 300 metres wide, locally even wider, marks the boundary between the Telemark formation and the Kongsberg formation. Along the breccia great upthrusts have occurred, bringing the Kongsberg formation, which is presumably the older, on a level with the younger Telemark formation". Starmer (1977, 1979) has shown that a more fundamental division exists in the west, where a mylonite zone, often exceeding 1km in width, trends in a N-S direction. The so-called 'friction breccia' is a late brittle fracturing concentrated largely around the mylonite zone.

1.3 LITHOLOGIES OF THE KONGSBERG SERIES

1.3.1 The variable quartz-plagioclase-biotite gneisses

These supracrustals are suggested to represent mixed metavolcanics and metasediments (A. Bugge, 1937). They show all modal variations; hornblende and epidote minerals may be major constituents. Almandine and muscovite are also common. The gneisses to the west of Kongsberg are generally quartz-rich and, in the Grøsli area, are often bi-mineralic quartz-feldspar rocks. East of Kongsberg, the rocks are generally coarser with biotite and hornblende becoming more dominant. Rocks in the west are frequently cataclased and, in the extreme cases, mylonitized. Effects become weaker eastwards, away from the mylonite zone, with sporadic effects restricted to narrow zones (Starmer, 1977).

J. Bugge (1943) described the rocks as dacitic gneisses and alternating dark and light bands consisting of amphibolites, biotite quartzites, dioritic and granitic gneisses. He considered these rocks to be the oldest in the 'Konsberg formasjon'. These variable gneisses are described more fully in Chapters Three and Seven.

1.3.2 The hornblende gneisses

These rocks form major bands only to the south-west of Kongsberg. Elsewhere (including the Eiker area) these rocks occur as thin discontinuous bands in the variable gneisses. They are interpreted as basic volcanics and vary from plagioclase-rich amphibolites to hornblende-rich quartzose gneisses.

1.3.3 The 'Fahlbånds'

These are 'rust zones' occurring within the supracrustals. (This term originated from the old German miners word, 'Falle', meaning rust). When weathered these zones are very friable due to the decomposition of sulphide minerals (mainly disseminated pyrite) present in the rock. They occur over the whole of the Kongsberg Series, but are noticeably absent in the Grøsli region. At several localities around Kongsberg, the fahlbånds may have acted as 'fixers' of metallic elements present in the basic 'Vinor' magmas.

1.3.4 The amphibolites and dioritic gneisses

The dioritic gneisses appear to be restricted to the south-west of the Kongsberg Series. They are equivalent to the 'plutonics' of J. Bugge (1943) who described them as gabbro-diorites, quartz-hornblende diorites and quartz-biotite diorites. He suggested that they may be 'dioritic differentiates' of basic magmas. West of Kongsberg, the dioritic gneisses contain no major amphibolite bands. Their mineralogy is plagioclase-hornblende-quartz with biotite, almandine and epidote often present. Although similar in mineralogy to the hornblende gneisses, they are much coarser and less foliated except where cataclased. East of Kongsberg, amphibolites dominate over dioritic gneisses but the two rock types are often associated in the same bodies, suggesting a genetic affinity. Margins between the two types can be either sharp or gradational.

1.3.5 The 'Vinor' basic intrusives

This group of intrusions were named the 'Vinordiabas' rocks by A. Bugge (1937). Their petrology is similar to that of the 'hyperites' of the Bamble Series along the Skagerrak coast. They comprise a number of basic injections which formed stocks bosses and dykes, during, and after, the waning stages of the major cataclasis that occurred in the west of the Kongsberg Series. They have been amphibolitized to varying degrees and are intruded into the supracrustals with the largest bodies occurring in the area around Grøsli. The Grøsli mine is situated on the margin of one such large body, where a sequence of injection phases consisted of early olivine gabbros and gabbros, with subsequent more hydrous gabbros and very late, fine grained amphibolite dykes.

A feature which occurs extensively in the Grøsli area, but rarely elsewhere in the Kongsberg Series, is the mobilization of the acidic supracrustal gneisses; these have intruded the basic rocks, forming veins and stock-works. The latest 'Vinor' intrusives (a series of dykes) cross-cut the mobilized gneisses and intrude the supracrustals both concordantly and discordantly. In the Eiker district, all of the intrusive basic bodies are thoroughly amphibolitized.

1.4 STRUCTURE OF THE KONGSBERG SERIES

The foliation and litho-banding of the Kongsberg Series has a predominant N-S strike and vertical to steep easterly dip. In the western part of the Series, there are no major folds. Minor isoclinal folds with axial planes parallel to the regional foliation are present and may be due to the movements which produced the cataclasis. Other folds, outside the cataclased areas, do not have axial planes in the foliation and may be earlier. The pre- and syn-cataclasis fabrics have been folded by later, open concentric folds which also affected the 'Vinor' dykes. Supracrustals tended to be bent around the relatively competent, larger 'Vinor' bodies; this effect is well seen at both the Grøsli and Eiker mines (see Chapters Two and Six).

1.5 SHEARING AND METAMORPHISM OF THE KONGSBERG SERIES

Three phases of shearing have been recognized (Starmer, 1977). The earliest and most severe was the cataclasis that formed the mylonite zone in the west and diminished eastwards into less strongly sheared rocks in narrow zones. Cataclasis occurred before and in the early stages of the Sveconorwegian orogeny. After the climax of this orogeny, lower grade metamorphism heavily overprinted the western rocks with Epidote-Amphibolite and Upper Greenschist facies assemblages. However, eastwards this effect diminished and the rocks retained their higher grade assemblages (e.g. the sillimanite and almandine-hornblende bearing rocks at Eiker).

A second period of movement produced the 'friction breccia'. Movements were again concentrated in the west and occurred under Mid to Lower Greenschist facies conditions. They were brittle movements and accompanied by little or no recrystallization. The brittle fractures in the shear zone at Eiker may be of this age. Finally, late Precambrian fractures developed at Lower Greenschist facies grade along well defined N-S fault planes. A number of Phanerozoic movements formed new faults and re-activated earlier shears.

1.6 THE OSLO GRABEN

Cambro-Silurian sediments of the Oslo Graben cover the eastern margin of the Kongsberg Series (Fig. 1.1). A Bugge (1937) considered that the junction was faulted in the south. Although there are many faults near the junction of the Oslofelt, the sediments actually lie unconformably on the Precambrian Kongsberg Series.

The Oslo Graben was tilted about 15° to the south-east in Phanerozoic times (Starmer 1977). This resulted in re-activation of some Precambrian faults, the formation of new faults parallel to the Oslo Graben margin and the development of concomitant E-W trending conjugate fractures. Two kilometres to the north of the Eiker shear zone (which trends N.E.-S.W.), is a fault zone trending E-W which was mineralized in the Permian. These two shears may represent conjugate faults which suggests movement of the Eiker shear during the Phanerozoic.

Hydrothermal veining, linked with Permian activity of the Oslofelt, occurred extensively in the Kongsberg Series. Quartz-calcite-fluorspar veins often trend approximately E-W. Where they have crossed 'fahlbånds', native Ag is associated with Cu, Zn, Fe, Ag and Pb sulphide mineralization in many workable deposits.

The geological history of the Kongsberg Series has been summarized by Starmer (1977) and has been reproduced in a modified form in Table 1.1. Although there are differences in the lithologies from east to west in the Kongsberg Series, the general metamorphic history is the same throughout the region.

TABLE 1.1

SUMMARY OF EVENTS IN THE KONGSBERG SERIES

(Modified after Starmer, 1977)

<table>
<tr><th colspan="3">EVENTS</th><th>METAMORPHISM</th></tr>
<tr><td colspan="3">SUPRACRUSTALS: Variable quartz-plagioclase-biotite gneisses. Hornblende gneisses.</td><td rowspan="3"></td></tr>
<tr><td colspan="3">DEFORMATION: Minor isoclinal folds.</td></tr>
<tr><td colspan="3">INTRUSIONS: Now forming amphibolites and dioritic gneisses.</td></tr>
<tr><td colspan="3">FOLDING: Regional fabric developed; minor isoclines with axial planes in foliation.</td><td rowspan="2">SVECOFENNIAN
Upper Amphibolite facies.</td></tr>
<tr><td colspan="3">GRANITIZATION: Forming granitic rocks east of Kongsberg.</td></tr>
<tr><td rowspan="2">GRANITES west of Kongsberg</td><td rowspan="2">BASIC 'VINOR' INTRUSIONS gneiss mobilization</td><td>CATACLASIS and syntectonic recrystallization</td><td rowspan="2">SVECONORWEGIAN
Mid-Amphibolite facies.</td></tr>
<tr><td>Post-tectonic recrystallization.</td></tr>
<tr><td colspan="2"></td><td>Severe overprinting in west</td><td>Epidote-Amphibolite facies.</td></tr>
<tr><td colspan="3">GRANITES: West of Kongsberg</td><td rowspan="2">Upper Greenschist facies</td></tr>
<tr><td colspan="3">FOLDING: Open concentric folds</td></tr>
<tr><td colspan="3">'FRICTION BRECCIA'</td><td>Mid-Lower Greenschist</td></tr>
<tr><td colspan="3">Late Precambrian faults</td><td>Lower Greenschist</td></tr>
<tr><td colspan="3">Phanerozoic sediments</td><td rowspan="3"></td></tr>
<tr><td colspan="3">Tilting and formation of Oslo Graben</td></tr>
<tr><td colspan="3">Permian intrusives and hydrothermal activity.</td></tr>
</table>

CHAPTER TWO

LITHOLOGIES, STRUCTURE AND AGE RELATIONSHIPS AT GRØSLI

2.1 THE GABBRO-SUPRACRUSTAL CONTACT

The Grøsli mine is situated in the west of the Kongsberg Series, five kilometres east of the junction with the Telemark Series (Fig. 1.2) from which it is separated by a major mylonite zone. Sulphide mineralization occurs across the contact of coarse-grained, leucocratic, supracrustal gneisses with coarse-grained gabbros and foliated/non-foliated amphibolites (the 'Vinor' intrusives).

Twelve boreholes were drilled, their location being based on the data from a very low frequency (V.L.F.) survey (Fig. 2.2). Fig. 2.3 has been drawn to show the borehole data as clearly as possible. A surface geological map (Fig. 2.1) and isometric block diagram (Fig. 2.4) show the relative positions of the boreholes and their relationship to the surface geology. As shown in Fig. 2.3 the gabbro-supracrustal contact is complex with much interpenetration. The contacts trend approximately north-south in all cases. Fig. 2.4 shows a continuous series of north-south apophyses from the roof of the gabbro, with intervening gneisses and the 'Vinors' sometimes breaking through the present erosion surface. It is emphasised that these do not represent folds but elongate intrusive shapes with long axes north-south.

An alternative interpretation of the intrusion shapes is shown for boreholes 1, 7 and 10. Berthelsen (1970) suggested the term 'globulith' for such lobate intrusions of basic rock into the plastic gneisses of the Moss area, south-east Norway. Starmer (1979) has noted these structures in a few places in the northern and western continuations of the large Vinor gabbro body at Grøsli. In all exposures, the strike and dip of the gneisses is sub-concordant to that of the 'Vinor' bodies. Rare, thin, amphibolite veins cross-cut the gneiss foliation for 10-20 cm from the contact and occasional xenoliths of gneiss are present within the 'Vinor' bodies.

Junctions are sharp and of two types. Some are an interfingering of 'Vinor' and gneiss bands (5 to 50 cm wide) concordant with the foliation of the gneisses. Other junctions consist of angular blocks of gabbro separated by veins of mobilized gneiss material (Fig. 2.5). In the latter, gneisses have been mobilized by the increased heat flow and fluid-vapour activity associated with the intruding 'Vinor' bodies, which they have back-veined.

2.2 THE SUPRACRUSTALS

These are dominantly leucocratic, coarse-grained rocks which have a general dip of 60^{o} east to vertical with rare westerly dips. The rocks are poorly to well-foliated, depending on the amount of biotite present. In the immediate vicinity of the Grøsli mine, quartz rich rocks dominate to the east of the gabbro contact. In this zone, amphibolite bands, which are obviously part of the supracrustal sequence, are extremely rare and always less than 5 cm thick. Further east, these bands become thicker and more numerous. Starmer (1977) considered the foliations in the gneisses of the Kongsberg Series to be transposition structures, therefore masking the original stratigraphic banding. Transposition could well result in the present foliations being at

FIG. 2.5. Back veining of gabbro by mobilized gneiss.

FIG. 2.6. Metagabbro dykes intruded into coronite.

high angles to the original bedding direction (Hobbs, Means and Williams, 1976). For this reason, apparent isoclinal fold axes have not been included on Fig. 2.1.

Granitization and microcline porphyroblastesis, common features in the Telemark Series, are rare within the Kongsberg Series gneisses and absent in the gabbro body, which apparently intruded after these processes. Gneisses now close to the gabbro bodies have a higher biotite content than elsewhere, possibly due to the diffusion of iron out of the basic body. The biotites formed in this manner can be distinguished from supracrustal biotite by their greater size and more random orientation.

Large open folds and flexures were produced after the consolidation of the gabbro bodies and developed concordant foliations in gneisses adjacent to gabbro outcrops. This effect is due to the fact that the gabbros presented a competent mass around which the more plastic gneisses could be moulded. To the east, away from the 'Vinor' gabbro bodies, . deformations of the same age have produced large amplitude folds with gently dipping axes.

2.3 THE 'VINOR' GABBROIC INTRUSIVES

Three separate phases of intrusion can be identified, the first stage constituting the majority of the 'Vinor' body around the Grøsli mine. This earliest stage is now represented by coronite, exposed in the centre of the basic bodies (Figs. 2.1 and 2.3). More water rich dykes intruded the earlier gabbro and now exhibit both sharp and diffuse contacts (Fig. 2.6). These dykes vary from heavily serpentinized gabbros to metagabbros. The formation of most of the metagabbro in the area however, is due to the metamorphism of coronite. In both surface and borehole geology, there is a gradation outwards in the body from coronite to metagabbro and finally amphibolite, indicating progressive metamorphism. Very often, the amphibolite at the junction with the gneisses has a very strong foliation which becomes less intense towards the interior. This amphibolite is often biotite rich as a result of the diffusion of potassium from the gneisses. In a few localities, garnet-biotite-chlorite and actinolite-chlorite schists have been produced at the 'Vinor' margin(for a maximum width of two metres)and are due to later (Greenschist grade) movements along the contacts.

Contamination by mobilized gneisses has led to the production of more felsic or dioritic rock, at the margin, and sometimes persisting for several metres inwards (Fig. 2.7). These rocks have not been represented on the maps since they are highly irregular in development and of low volume. The latest stage of basic intrusion is represented by small, fine grained amphibolite dykes with sharp margins. They are rare and have not been found at the surface but occur in a few borehole sections. These dykes cross-cut the previous two intrusive phases, their metamorphic derivatives and the mobilized gneiss veins. They also cross-cut massive sulphides (which have not been mobilized by later fluids) within the 'Vinor' gabbro body, and can be seen to carry trails of ore fragments. These dykes are, however, traversed by several metamorphic segregational features. In both amphibolite and metagabbro bodies, zones of sodic plagioclase and actinolite rock were sweated-out of the parent. In some places, these segregations form ptygmas, but

FIG. 2.7. Mobilized gneiss vein and dioritic metagabbro.

FIG. 2.8. Segregational pegmatite veins intruded into metagabbro.

also occur as veins which cross-cut the rocks with both sharp and diffuse boundaries. At the surface they form pegmatitic veins from 1 cm to 1 m across. At least two phases of these metamorphic segregations are discernible in the field (Fig. 2.8). A small scale feature of the 'Vinor' rocks is the presence of large garnet porphyroblasts which have grown over the foliations of some of the amphibolites.

The actual shape of the 'Vinor' gabbro bodies is difficult to determine from the borehole data. It is evident that the intrusion is of complex shape and not related to any fold geometry, as is to be expected since intrusion occurred into country rocks at elevated temperatures at the start of the Sveconorwegian metamorphism. In the Moss area of south-east Norway, gabbros intruding heated country rocks were called 'globuliths' by Berthelsen (1970). He defined the term as "an intrusive body or a group of closely associated bodies of globular or botryoidal shape and with almost concordant contacts resulting from the effects of the intrusion(s) on its/their immediate surroundings". He assumed that the gneisses were pre-heated under high water vapour pressure so that they were close to the conditions of anatexis. This assumption was made because even small bodies had anatexis at their margins; considered to be due to the heat given off by the solidifying magma. He stated that 'contact deformation' was caused in the supracrustals by stress differences directly due to the specific intrusive mechanism of the globuliths and that the plastic style of the structures may have been brought about by the softening of wall rocks by contact anatexis. Obvious similarities can be drawn with the Grøsli area structures although there are also major differences. Partial anatexis has occurred, but only on a minor scale and with a body as large as the

Grøsli gabbro, a great deal more gneiss mobilization would be expected with Berthelsen's model. Water was definitely not present in large amounts to promote anatexis or to cause swelling of the magma to form globulithic shapes. In fact, a paucity of water is indicated by the presence of quite large volumes of coronite still preserved. Berthelsen himself stated that there is no definite method of identifying the concordant structure of the gneisses as 'contact deformation' features rather than later superimposed foliation.

2.4 THE ORE DEPOSIT

The pyrrhotite-pyrite-sphalerite-chalcopyrite ore occurs either as massive deposits or as impregnations in both gneisses and 'Vinor' intrusives. The only large deposit exposed at the surface is in the north of the area, at the disused Grøsli mine (Fig. 2.9). The ore body is located on the top of a rounded Vinor gabbro, mainly within the gneisses but also impregnating the gabbro itself.

There is no definite axis of ore concentration. This is to be expected in rocks which were undergoing high-grade regional metamorphism and behaved differently to those found in higher level tectonic environments (e.g. the Caledonides). The zone of mineralization becomes deeper southwards and since there are no recorded ore outcrops to the north of the mine there may be a very broad 'axis' of ore deposition dipping southwards at 20° to 30°. In this case to the north of the mine, the 'axis' would be above the level of erosion and would be at progressively greater depths southwards. This zone is not continuous in ore grade along its length. The massive ore bodies (60-100% modal ore) are

FIG. 2.9. Massive ore exposed at the surface in the disused Grøsli mine.

FIG. 2.10. Shear zone in wall of Grøsli mine.

discontinuous both horizontally and vertically and appear to be irregularly lensoid or more roughly spherical in shape. In the mine, one feature of the hetrogeneity is the encrusting of host rock xenoliths producing what appears to be a massive outcrop, but which is in fact a thin veneer only 2-3 cm thick. Hetrogeneity of the ore, and the depth to which the zone appears to plunge, would cause severe problems in mining. The only other evidence of mineralization at the surface is gossan or 'hardpan' in the south of the area (Fig. 2.1) with an outcrop area of less than 1 m^2.

Host rock alteration associated with the ore bodies is variable and may be absent. Where present, the adjacent ore body is either within the gneisses or situated close to their junction with the gabbro bodies. Ore bodies deeper within the gabbros have no wall rock alteration. The alteration minerals are dominated by chlorite with subordinate white micas and accessory calcite, magnetite, quartz and spinel. The altered rock is pale green or grey-green in hand specimen and the original parent is often difficult to identify although the presence of substantial quantities of primary quartz and garnet are indicative of a gneiss. The alteration is very localized and nowhere does it extend more than ten metres from a massive ore body, even when dissemination is present in the host rocks. This fact will be considered later in discussion of relative age relations between ore and wall rock alteration (Chapters Three and Eleven). A small number of quartz veins containing sulphides cross-cut the alteration zones as well as the ore bodies.

A late brittle shearing event, localized in the north of the area, is recorded only at the mine. A shear (dipping 45^{o} east to sub-horizontally) reflects upthrusting of gneisses from the east over the ore body (Fig. 2.10). The displacement was several metres and a fault-breccia up to 1 m wide containing gneiss fragments and oxidised sulphide material has been formed. This shearing event is probably of the same age as the minor movements along the gabbro junction which produced the biotite-chlorite schists.

2.5 RADIOMETRIC DATING

O'Nions and Heier (1972) presented Rb-Sr isochrons for supracrustals of the Kongsberg area about 25 km south of the Grøsli mine area. They concluded that the supracrustals were metamorphosed at least twice; during the Svecofennian orogeny (1700 ± 100 m.y.B.P.) and during the lower grade Sveconorwegian orogeny (1260 ± 40 m.y.B.P.). Jacobsen and Heier (1978) determined Rb-Sr isochrons for some rocks from the Kongsberg series, including a sample of supracrustal quartz-plagioclase gneiss (12 km east of Grøsli) and five analyses from the 'Vinor' body associated with the Grøsli mine. They deduced an age of 1600-1500 m.y.B.P. for the Svecofennian orogeny and an age of 1200-1100 m.y.B.P. for the Sveconorwegian event. The intrusion ages of the gabbroic 'Vinor' bodies were placed between 1370 and 1070 m.y.B.P. with the main intrusions probably occurring from 1370 to 1200 m.y.B.P. The 'Vinors' thus appear to have been intruded prior to, or synchronously with the commencement of the Sveconorwegian orogeny.

2.6 METAMORPHISM OF THE 'VINOR' GABBROIC BODIES

From field evidence, several observations can be made concerning the history of the basic intrusives. The major intrusion was largely derived from an early injection phase (now forming coronites) with minor additions at later stages. In the major gabbro body, the effects of subsequent metamorphism decreased progressively inwards with a sequence of decreasing hydration (amphibolitization) indicating a limited access of water. A paucity of water would be expected since the supracrustals had undergone a previous high-grade metamorphic event and would therefore be essentially anhydrous. The initial metamorphism of the gabbros was probably fairly soon after, or perhaps synchronously with, their emplacement, during the Sveconorwegian orogeny.

CHAPTER THREE

PETROLOGY OF THE SILICATE ROCKS AT GRØSLI

3.1 THE CORONITES

These gabbroic rocks have retained their primary igneous mineralogy and textures, with the development of narrow coronas between olivine and plagioclase and around magnetite. The majority of the metagabbros and amphibolites of the mine area have developed from metamorphism of these intrusives. The rocks are all either gabbros or olivine gabbros, with clinopyroxene dominant over orthopyroxene. Textures are usually sub-ophitic (Fig. 3.7) and occasionally ophitic, but orthocumulate textures are also present, with olivine and plagioclase forming the cumulate and pyroxenes the intercumulate fractions. Olivines are often idiomorphic (up to 7mm long) and are usually partially enclosed by plagioclase laths (up to 9mm long), although olivine sometimes encloses plagioclase. Xenomorphic magnetite and ilmenite (up to 3mm across) usually enclose both olivines and plagioclases, whilst pyroxenes (up to 2cm^2) surround all other minerals. Pleonaste and apatite occur as accessory minerals.

3.1.1 Olivine

Four representative analyses of specimens ranging from hyalosiderite to hortonolite (Fa_{48-54}) are presented in Table 3.1. The olivines show all stages of progressive deuteric alteration. Often the crystals are unaffected except for cracking, which contains iron oxides. The more extensive replacements involve iddingsite, antigorite, talc and

TABLE 3.1

ELECTRON MICROPROBE MINERAL ANALYSES

	1	2	3	4	5	6
SiO_2	35.61	35.80	35.01	34.98	54.78	54.92
TiO_2	0.01	0.00	0.00	0.00	0.07	0.12
Al_2O_3	0.00	0.00	0.00	0.12	0.61	0.56
MgO	21.54	20.49	22.42	23.94	26.18	25.85
FeO	40.88	42.43	41.01	39.93	18.02	18.03
MnO	0.59	0.65	0.79	0.58	0.52	0.63
CaO	0.02	0.02	0.03	0.02	0.31	0.83
Na_2O	0.00	0.00	0.00	0.04	0.62	0.56
K_2O	0.00	0.00	0.00	0.02	0.00	0.00
Cr_2O_3	0.01	0.00	0.00	0.00	0.00	0.00
TOTAL	98.66	99.39	99.26	99.63	101.11	101.50

FeO = Total iron as FeO

Analysis 1 to 4. Olivines in coronitic gabbro.

5 and 6. Orthopyroxene coronas around olivines in coronitic gabbro.

carbonates. Some crystals are almost entirely replaced by iron oxides (Fig. 3.1). Coronas developed around the olivines are unaffected by this alteration and are a later phenomena (Fig. 3.2). Double coronas are extensively developed and single rims are uncommon. The ubiquitous innermost rim is narrow (0.05 - 0.1mm wide), and composed of orthopyroxene. Two microprobe analyses (Table 3.1) produced compositions of hypersthene (En_{59}). A second corona of green or green-brown amphibole is frequently present with a columnar crystal development perpendicular to the first rim: it is usually 0.1 - 0.2mm in width and larger than the inner corona. These outer rims often have vermicular or dendritic structures which have been previously described as a symplectic spinel-amphibole intergrowths (Starmer, 1969). Analyses of two such amphiboles showed hornblende compositions (Table 3.2).

A third and much less continuous rim, which is often associated with the olivines, is an intergrowth of vermicular sodium feldspar in a host of potassium feldspar (Fig. 3.3). Where the intergrowth is associated with the coronas it can occur instead of the outer amphibole rim or can be sandwiched between the two, indicating a contemporaneous formation with the other coronas.

Sederholm (1916) believed that coronites owed their origin to regional metamorphism. Mason (1967) also concluded that they are the product of a metamorphic reaction between olivine and the surrounding plagioclase. In some cases, late hydrothermal veins have cut the coronites and the fluids have totally altered the olivines to serpentinite and chlorite. In these cases, the coronas have also been affected.

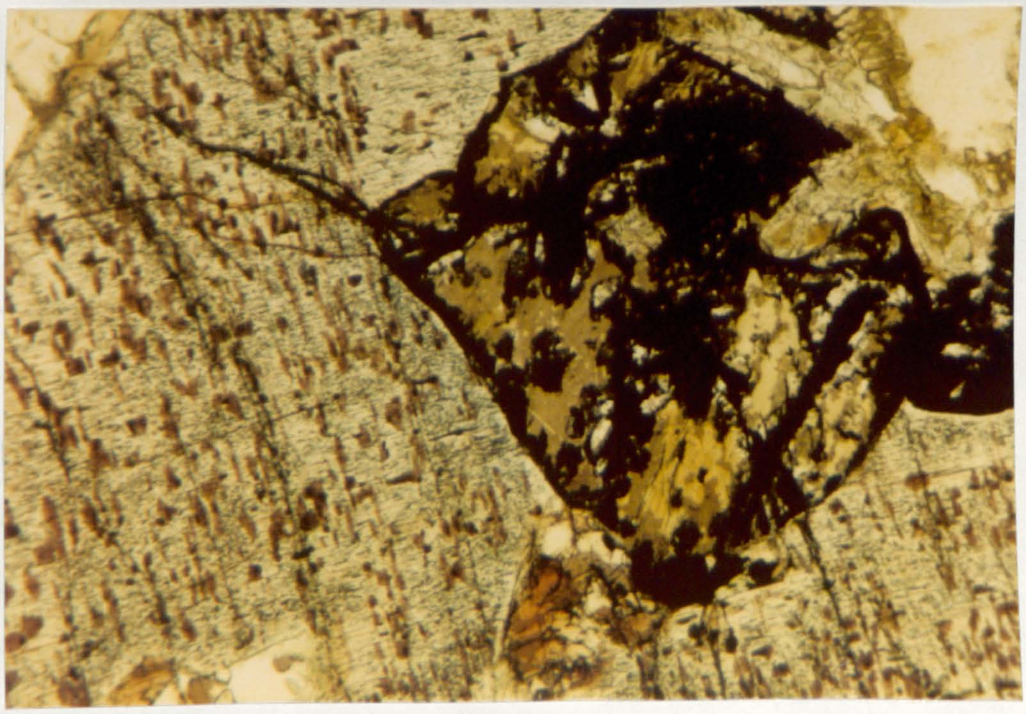


FIG. 3.1. Serpentinized olivine adjacent to unaltered clinopyroxene (containing exsolved rutile and magnetite). PPL. X12.5.

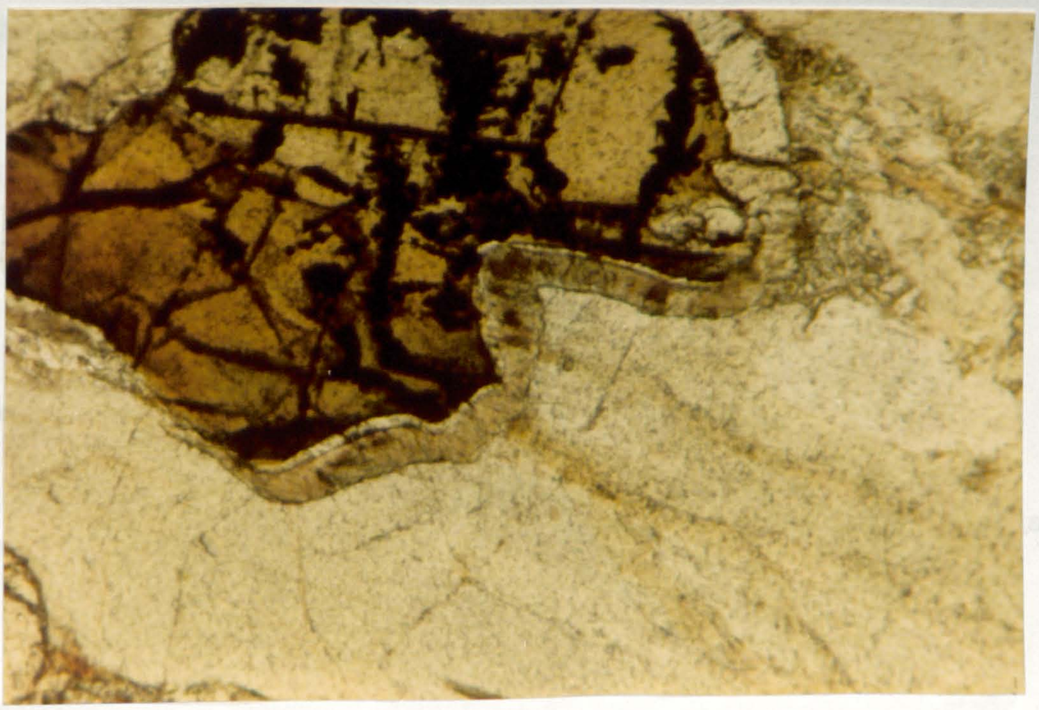


FIG. 3.2. Inner orthopyroxene and outer amphibole coronas developed around serpentinized olivine against plagioclase. PPL. X31.5.

TABLE 3.2

ELECTRON MICROPROBE MINERAL ANALYSES

	1	2	3	4	5	6
SiO_2	42.27	40.39	53.18	53.67	52.59	52.90
TiO_2	0.19	0.07	0.94	0.31	0.53	0.75
Al_2O_3	18.46	17.84	0.78	0.85	2.22	2.79
MgO	14.91	10.88	24.11	23.51	14.42	14.78
FeO	8.48	12.59	18.43	18.50	9.70	10.59
MnO	0.20	0.17	0.55	0.42	0.23	0.20
CaO	10.70	11.03	2.13	2.54	19.75	17.57
Na_2O	3.61	2.15	0.30	0.60	0.56	0.47
K_2O	0.98	1.07	0.00	0.01	0.01	0.08
Cr_2O_3	0.00	0.00	0.04	0.00	0.01	0.04
TOTAL	99.80	96.19	100.46	100.41	100.02	100.17

FeO = total iron as FeO

Analysis 1 and 2. Amphibole rim around olivines in coronitic gabbro.

3 to 6. Pyroxenes in coronitic gabbro.

TABLE 3.2 (contd.)

ELECTRON MICROPROBE MINERAL ANALYSES

	7	8	9	10	11	12
SiO_2	53.06	53.57	51.12	40.89	51.67	40.69
TiO_2	0.50	0.26	N.D.	N.D.	N.D.	N.D.
Al_2O_3	1.44	0.71	2.69	17.48	3.31	16.82
MgO	13.24	20.28	15.38	6.23	15.62	8.26
FeO	9.19	23.16	17.97	18.70	15.98	17.99
MnO	0.22	0.51	0.41	0.15	0.18	0.16
CaO	21.53	1.48	10.21	11.34	10.51	11.30
Na_2O	0.32	0.00	0.17	2.01	0.21	2.08
K_2O	0.03	0.01	0.04	1.24	0.07	0.52
Cr_2O_3	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	99.53	99.98	97.99	98.04	97.55	97.82

FeO = total iron as FeO

N.D. = not determined

Analysis 7 and 8. Pyroxenes in coronitic gabbro.

9 and 10) Core and rim amphibole pairs pseudomorphing
)
11 and 12) pyroxene in coronitic gabbro.

3.1.2 Plagioclase

Crystals occur as idiomorphic laths and exhibit polysynthetic twinning on the pericline and/or albite laws. Carlsbad twinning may be combined with either or both. The laths usually have normal zoning, but sometimes show reverse or oscillatory effects. Most of the laths are of labradorite composition but the maximum measured variation in one crystal was An_{66-37}. The laths show two types of clouding. The first type appears as a light dusting of opaques. Deer, Howie and Zussman (1963) state that this dusting is magnetite, ilmenite or hematite with silicates such as garnet, biotite and hornblende. Starmer (1969) noted that the clouding disappears on the more sodic edges of many zoned crystals in contact with coronas and considered that this was due to the resorption of the dust to form amphibole. The second type of clouding (Fig. 3.4) is much denser and is due to hydrous alteration forming sericite in plagioclase. Within the coronites, this alteration is not widespread and is generally restricted to the more calcic cores of the laths.

3.1.3 Pyroxenes

Primary clinopyroxenes dominate over orthopyroxenes in the coronites of the Grøsli area. The orthopyroxenes are clean crystals with few or no inclusions but the clinopyroxenes show varying degrees of clouding caused by the exsolution of rutile and magnetite along crystallographic planes (Fig. 3.1). Rarely, thin lamellae of a second clinopyroxene (pigeonite) have exsolved. Microprobe analyses of three orthopyroxenes and three clinopyroxenes gave hypersthene and augite compositions respectively (Table 3.2). The pyroxenes have often developed very thin

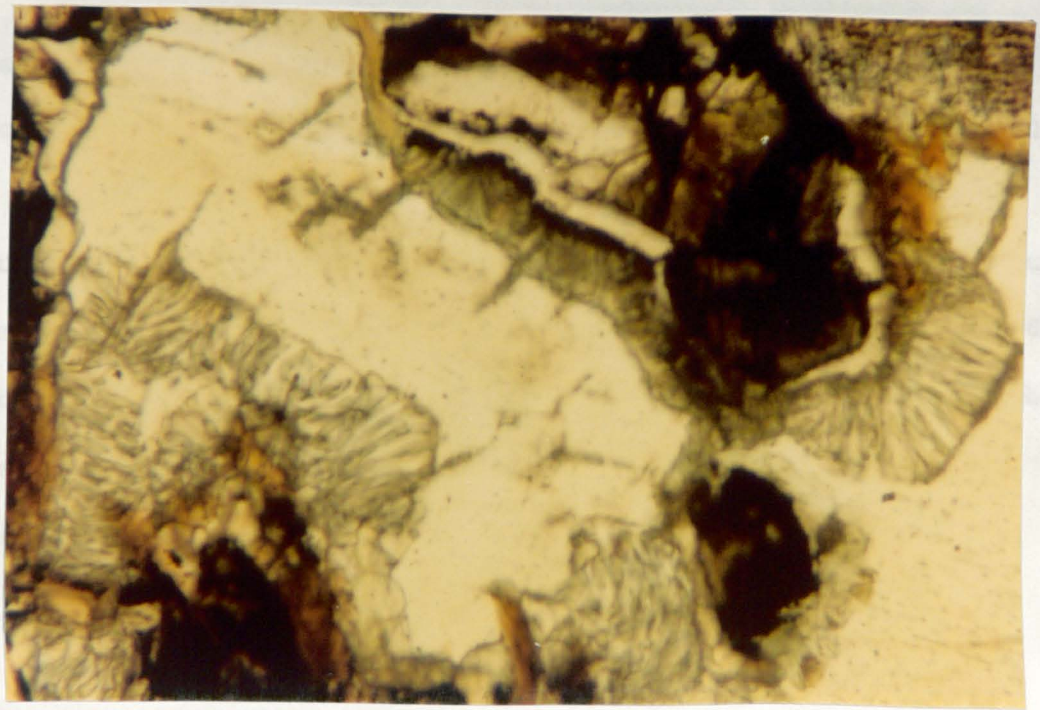


FIG. 3.3. Serpentinized olivines with a discontinuous corona of vermicular sodium feldspar in a host of potassium feldspar. PPL. X50.

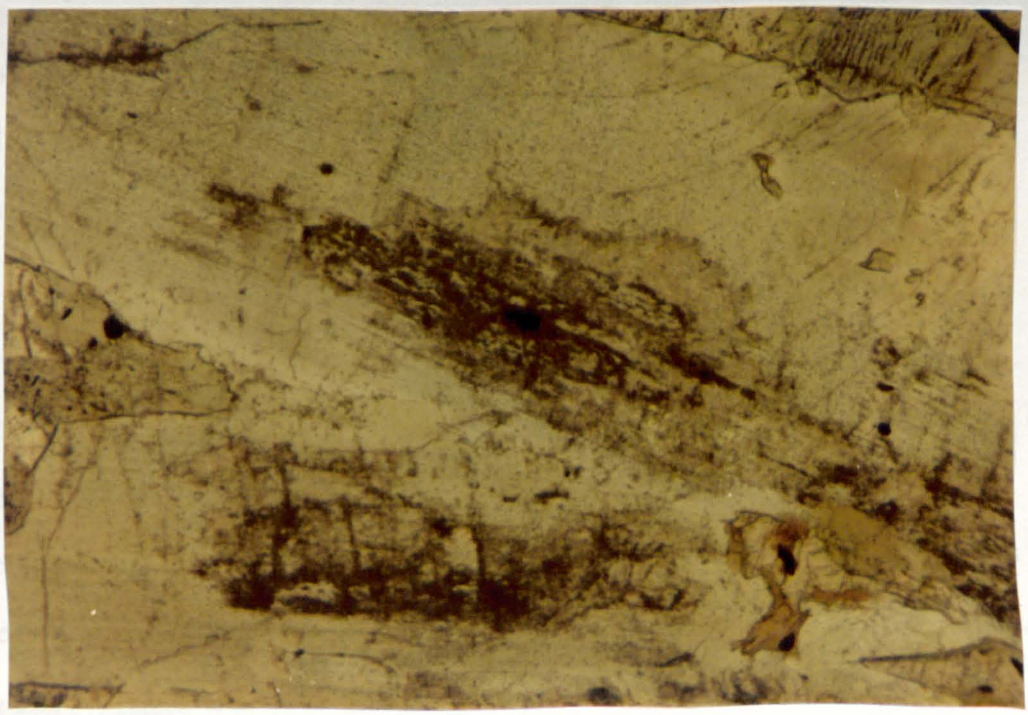


FIG. 3.4. Heavy dusting of plagioclase due to sericitic alteration. PPL. X12.5.

coronas (0.03mm thick) of green, green-brown or brown amphibole or of biotite (Fig. 3.5). These coronas are usually incomplete and only occur where pyroxene is adjacent to plagioclase. The rims are continuous around adjacent crystals of pyroxene and olivine and are thus contemporaneous with the outer corona of the olivines. Starmer (1969) stated that the thinner rim around the pyroxene was due to near stability of the pyroxene-plagioclase assemblage, where access of water was inhibited.

3.1.4 Magnetite

There are two modes of occurrence in the coronites, as primary magmatic precipitates and as alteration products within olivines (Fig. 3.6). Biotite forms coronas around magnetite (up to 0.3mm thick) and is usually red-brown with no distinct cleavage or crystal habit. A second corona of green amphibole is often present and both coronas may be incomplete. Magnetite often contains exsolution lamellae of ilmenite. More rarely, discrete ilmenite grains occur and contain exsolution blebs of rutile.

3.1.5 Sulphides

In coronites with sulphide concentrations exceeding accessory amounts, the original mineralogy generally shows signs of early amphibolitization. The olivines in many sulphide bearing coronites are totally replaced, but often the orthopyroxene rim can be identified in various stages of alteration, whilst the amphibole rim may be unaffected. Corona development occurred after the introduction of the bulk of the sulphides which themselves post-dated the crystallization

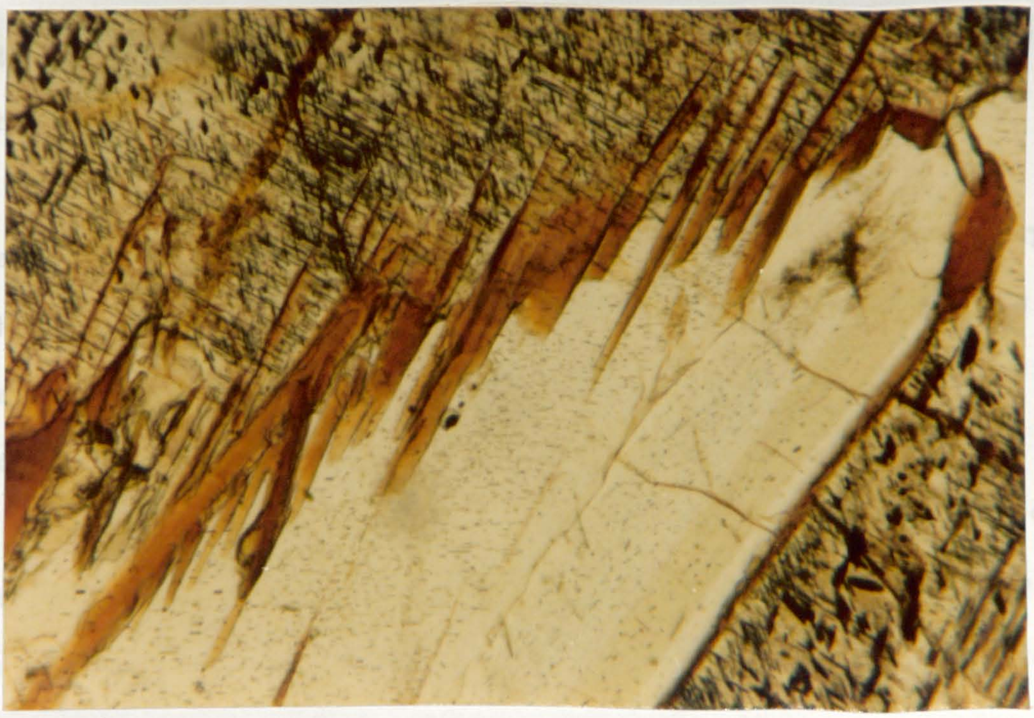


FIG. 3.5. Biotite corona around clinopyroxene adjacent to plagioclase. PPL. X50.

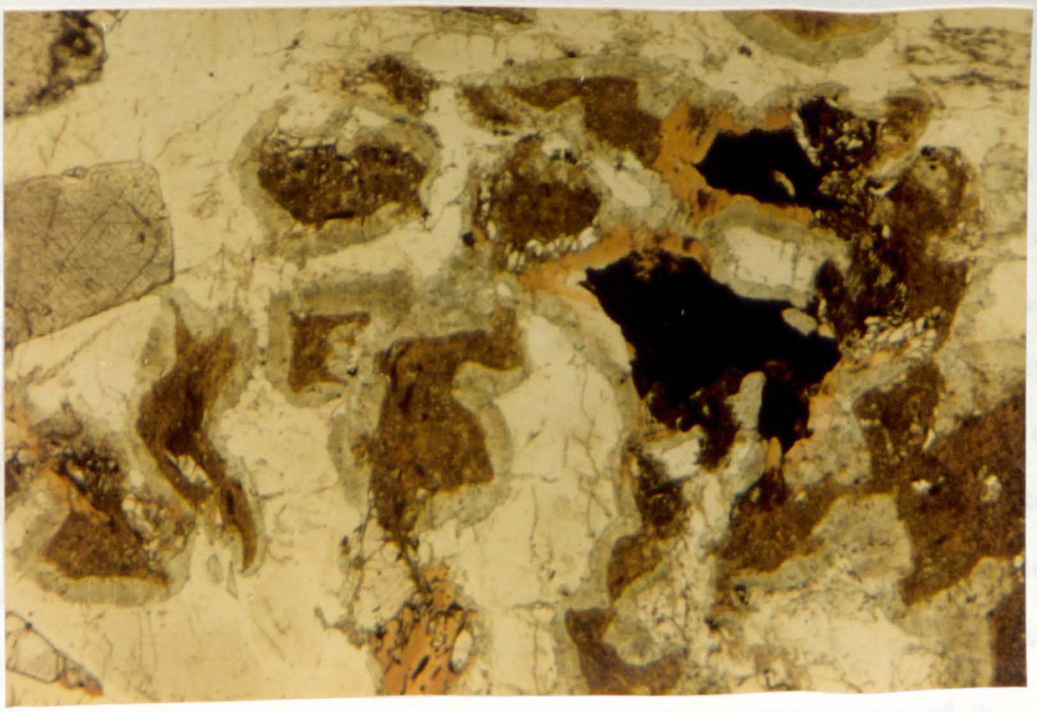


FIG. 3.6. Primary magmatic magnetite (with biotite coronas) and secondary magnetite (vermicular form with orthopyroxene rims) associated with altered olivine (brown on photo). PPL. X12.5.

of the early primary magmatic minerals. Magnetite is often veined by chalcopyrite and less frequently by pyrrhotite.

Sulphides also show extensive embayment and replacement of magnetite, penetrating along the octohedral cleavage directions. The sulphides commonly display double coronas: an inner rim of biotite and an outer rim of green amphibole. This outer rim is often continuous with that round adjacent olivine crystals.

Thus, the sulphide ore was introduced prior to the formation of the coronas and it is probable that corona development and amphibolitization occurred very shortly after the main sulphidization. However, later introduction of minor volumes of sulphides (mainly in veins with gangue calcite, chlorite, sericite and quartz) was linked to weak, low grade (Greenschist facies) metamorphism of the coronites.

3.2 THE METAGABBROS

This term is used to describe rocks which still retain igneous textures, but have undergone amphibolitization so that replacement of olivines and pyroxenes is almost complete or total. There is a continuous progression from coronites through metagabbros to amphibolites which have lost all relict igneous texture. The metagabbros then, are usually transitional assemblages sometimes containing primary pyroxenes etc. which are in the process of alteration to amphiboles. A second generation of metagabbros is present, but is only easily distinguishable where it intrudes coronites as thin dykes. The magmas carried a higher water content than those producing the gabbros which developed corona growths. Initially pyroxenes and plagioclases crystallized, but the high water content under the prevailing metamorphic conditions partially altered the pyroxenes to actinolite. This was thus an 'auto-metamorphism' since water was not introduced from outside.

The alteration of both generations of intrusives has produced a continuous variation to hornblende-oligoclase-epidote amphibolite and hornblende-labradorite amphibolite. All stages of the alteration can be recognized in the Grøsli area and often within a single thin section, underlining the variability in the amphibolitization. The alteration does show a general progressive increase towards the contacts with the gneiss, but no sharp divisions exist between coronite in the core of the bodies and amphibolite on the periphery.

3.2.1 Pyroxenes

The first detectable changes in the pyroxenes were represented by the development of a light green pleochroism. At this stage the crystals were heavily dusted and schillerized. The original crystal outline was still preserved. As alteration continued a patchy development of actinolite crystals occurred within the crystal (Fig. 3.8). This process continued with development of increasing pleochroism. The edges of the crystal became irregular as the primary shape disappeared. The final product of this early stage was an aggregate or patchwork of small actinolite crystals pseudomorphing the original pyroxene. Often in these metagabbros, a blue-green or green hornblende rim developed around the actinolite aggregates. The origin of these rims is discussed later, but in metagabbros which have lost all primary pyroxenes, these rims remain as ghosts and delimit the original pyroxene boundaries (Fig. 3.9). Table 3.2 shows

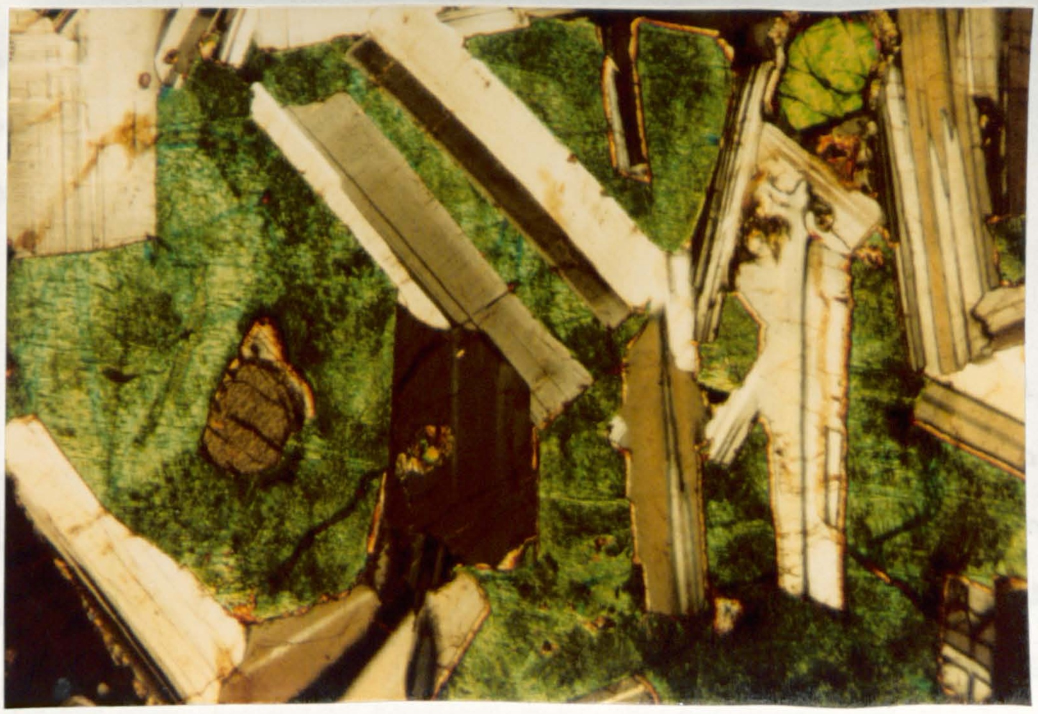


FIG. 3.7. Ophitic texture in coronitic gabbro (plagioclase and clinopyroxene). XPL. X12.5.

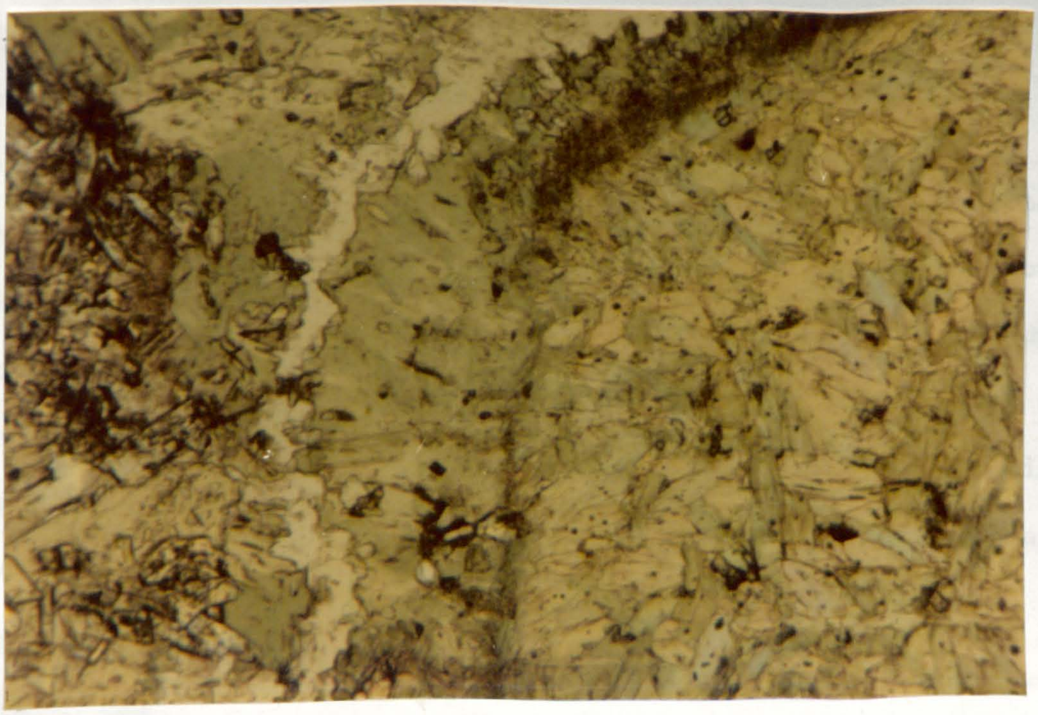


FIG. 3.8. Actinolite pseudomorphing pyroxene (right side of photo) with outer rim of chlorite. PPL. X31.5.

analyses of the pseudomorphed pyroxene cores which have actinolite compositions and of the rims which have more aluminous hornblende compositions.

3.2.2 Plagioclase

The transitional nature of the metagabbros is exhibited well by the plagioclase compositions. As with the pyroxenes, alteration of the original calcic plagioclases has proceeded to differing degrees. The coronites still retain the original feldspars (calcic labradorite to calcic andesine) and primary textures (Fig. 3.7). In more highly altered rocks, the plagioclases have recrystallized to short laths of sodic labradorite or andesine. With increased re-crystallization at this grade, the metagabbros have lost their primary textures and become amphibolites. However, where the rock has undergone subsequent epidote-amphibolite grade metamorphism, the plagioclases have been involved in several reactions, commencing with the formation of an epidote rim between the laths and the amphibolitized pyroxenes. Progressive epidotization (saussuritization) of the feldspars and/or reaction with other minerals produced a heavy overprint of epidote on the relict plagioclases. The result was an aggregate of epidote crystals surrounded by oligoclase and albite rims (Fig. 3.10). Chlorite is often present in these zones forming a rim to the amphibolitized pyroxenes (Fig. 3.8). In greatly altered specimens, the original lath shapes are obliterated due to reactions at their margins. In the metagabbros, discrete crystals of the more sodic plagioclases are uncommon. They occur as xenoblastic forms within the epidote-chlorite-albite/oligoclase association.

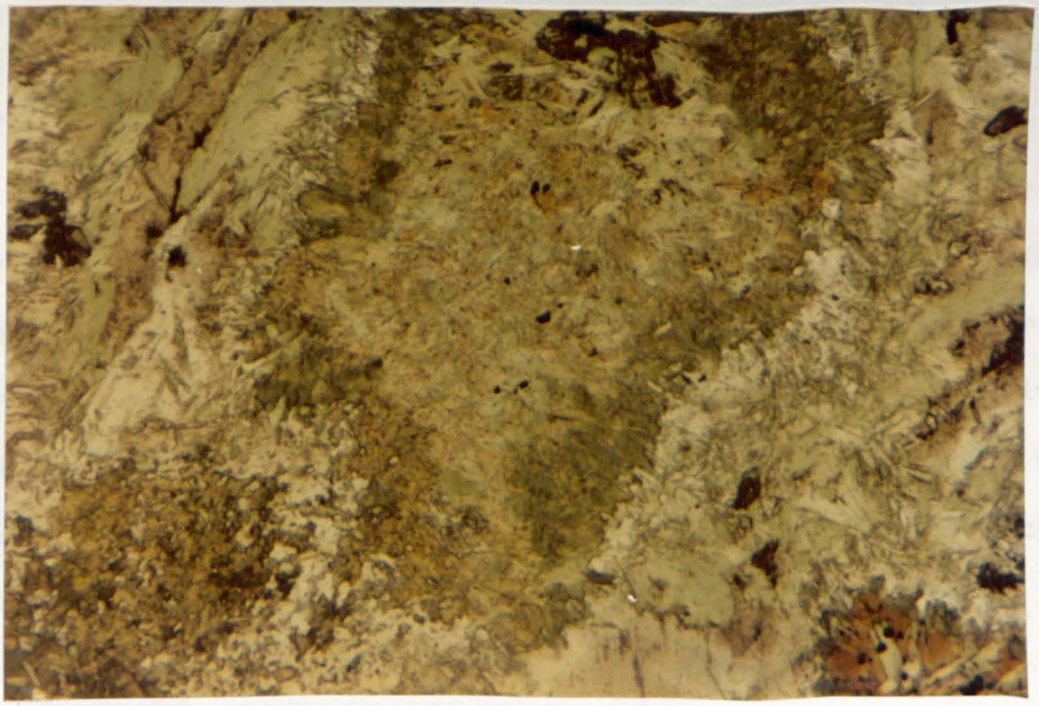


FIG. 3.9. Pseudomorph of pyroxene with actinolite core and hornblende rim maintaining original crystal outline. PPL. X12.5.

FIG. 3.10. Heavy saussuritization of original calcic plagioclase. Epidote minerals now occupy the core area with sodic feldspar at the margin. XPL. X31.5.

3.2.3 Epidote

Epidotes are not present in all metagabbros, but when they do occur, they are always associated with the primary plagioclase relics. In part they develop by the reaction of the anorthite component under hydrous conditions during the metamorphism. They may also be derived from the sericitic alteration products of the plagioclase, rather than directly from the feldspar. In some cases, the mineral forms at the junction of plagioclase and pyroxene crystals, through a reaction between the two. The lack of hornblendic rims on pyroxenes, which have this epidote border, suggests that the plagioclase may have reacted and resorbed this hornblende to produce the epidote.

3.2.4 Chlorite and biotite

Metamorphic biotite is not common in the metagabbros (as it is in the amphibolites). It occurs as clusters of xenomorphic plates often replacing amphibole (itself after pyroxene) and also as remnant coronas around ores. Biotite may also occur as larger ragged laths which have no obvious replacive nature. They are red-brown, brown or, rarely, green in colour. Katada (1965) states that the green biotites are formed at conditions below the amphibolite facies while the distinctly red-brown biotites (often rimming the magnetites in the coronites) are of at least mid-amphibolite facies grade. The biotites may or may not contain zircons with pleochroic haloes. Often the biotites show strained extinction and are in the process of altering from amphiboles, or hydrating to chlorites. Table 3.3 shows microprobe analyses of three brown biotites which lie in the middle of the common biotite field of Deer, Howie and Zussman (1963).

TABLE 3.3

ELECTRON MICROPROBE MINERAL ANALYSES

	1	2	3	4	5	6
SiO_2	35.81	35.88	35.09	25.19	27.00	27.34
TiO_2	3.14	3.15	3.04	0.10	0.06	0.06
Al_2O_3	15.70	16.00	16.46	21.15	20.52	20.43
MgO	10.40	10.53	10.35	14.05	13.73	13.91
FeO	21.96	22.04	21.56	26.22	25.42	25.32
MnO	0.35	0.38	0.35	0.62	0.64	0.63
CaO	0.01	0.06	0.03	0.02	0.02	0.03
Na_2O	0.07	0.05	0.04	0.00	0.00	0.00
K_2O	9.04	8.93	9.44	0.01	0.00	0.00
Cr_2O_3	0.02	0.03	0.05	0.02	0.00	0.03
TOTAL	96.50	97.05	96.41	87.38	87.39	87.75

FeO = total iron as FeO

Analysis 1 to 3 Biotites in metagabbro.

4 ot 6 Chlorites in metagabbro.

Chlorite is of very variable occurrence in the metagabbros; it can be absent or replace most of the mafic minerals present (usually in association with the opaques). Its habit is similar to that of the biotites: either as mats of small crystals or as large ragged plates. It also occurs as replacements of amphiboles and biotites. Three analyses are given in Table 3.3 indicating that the chlorites are of ripidolite composition.

3.2.5 Accessory minerals

Calcite is rare in the metagabbros. Irregular patches occur associated with chlorites, epidotes and altered plagioclases. It also occurs as late veins associated with the sulphides. Quartz is rare, becoming more common as the metagabbro is amphibolitized. It either occurs as small clusters of crystals showing strained extinction and lobate margins, or as later veins. Patches of very coarse grained felsic segregations are often present. They comprise zoned sodic feldspar laths (up to 10 x 5mm), often granulated at their margins (to 0.2mm size) and quartz. Other minerals such as secondary hornblende and biotite may occur in small quantities. The margins of all the crystals in these patches are either serrated or ragged, suggesting that they have been subjected to slight movements during their crystallization.

3.2.6 Opaques

Magnetite and sulphides occur frequently in all types of metagabbros. Magnetite never reaches major concentrations, unlike the sulphides. It is however ubiquitous in small concentrations and is a primary magmatic mineral: textural relationships suggest a later origin for the sulphides. Magnetite grains (1 - 5mm across) display different forms in the metagabbros and in the coronites. In the latter, the forms are sub-idiomorphic or interstitial and the maximum extent of reaction is the production of a biotite corona. In the metagabbros the magnetites have reacted and produced embayed and corroded outlines, the extent of reaction being related to the degree of alteration of the rock. The margins of the magnetite grains have become serrated and a thin border of sphene usually developed as the original igneous composition exsolved titanium (Fig. 3.11). The sphene is often closely associated with amphiboles, which may well have derived some iron from the magnetite. In extreme cases, reaction has left only skeletal magnetite with parallel rods defining the cleavage (Fig. 3.12).

Sulphides in the metagabbros do not show the complete interstitial characteristics of those in the coronites: they enclose and replace magnetite. They are extremely irregular in outline and enclose (and are enclosed by) many of the metamorphic minerals suggesting the presence of the ores within the rocks during metamorphism. Frequently, the sulphides are enclosed within ragged amphibole crystals, or mats, but are never present in the primary (igneous) silicates. Biotite is also commonly associated with the amphibole-sulphide association and may also be derived in part from the magnetite. The co-existence of amphiboles derived from primary pyroxenes and amphiboles with no such association, is explained by the origin of the latter from magnetite. Often the sulphides retain the form of a diffuse veining or network through the metagabbros and irregular bodies surround (and occasionally intrude) pyroxenes and plagioclases which are in the process of alteration

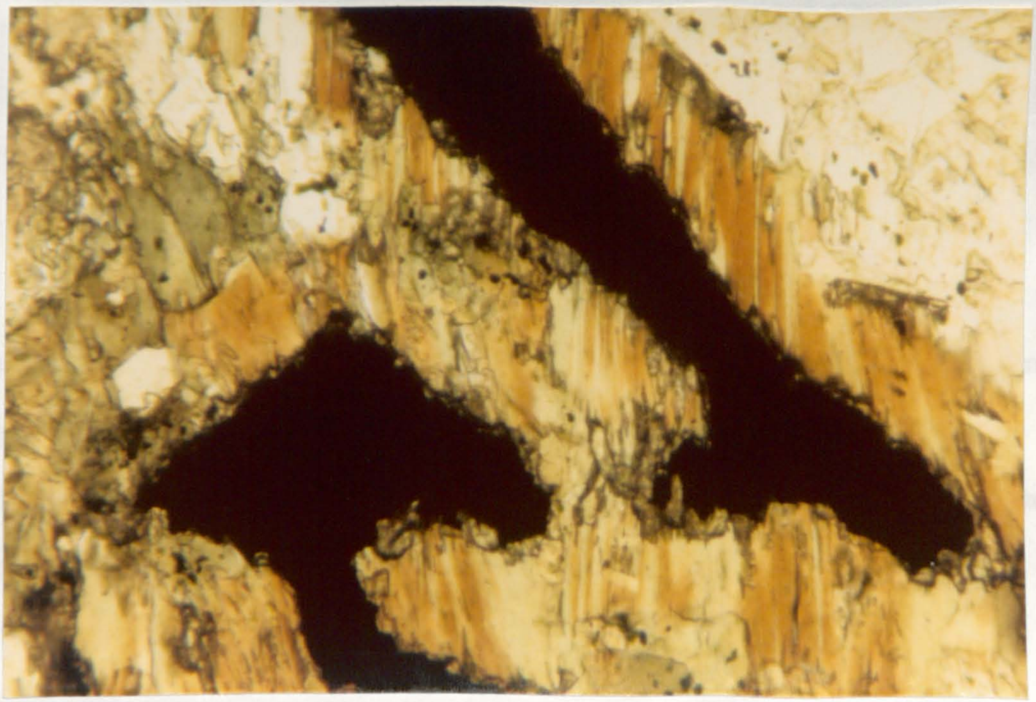


FIG. 3.11. Magnetite with rim of sphene surrounded by biotite in the process of chloritization. PPL. X31.5.

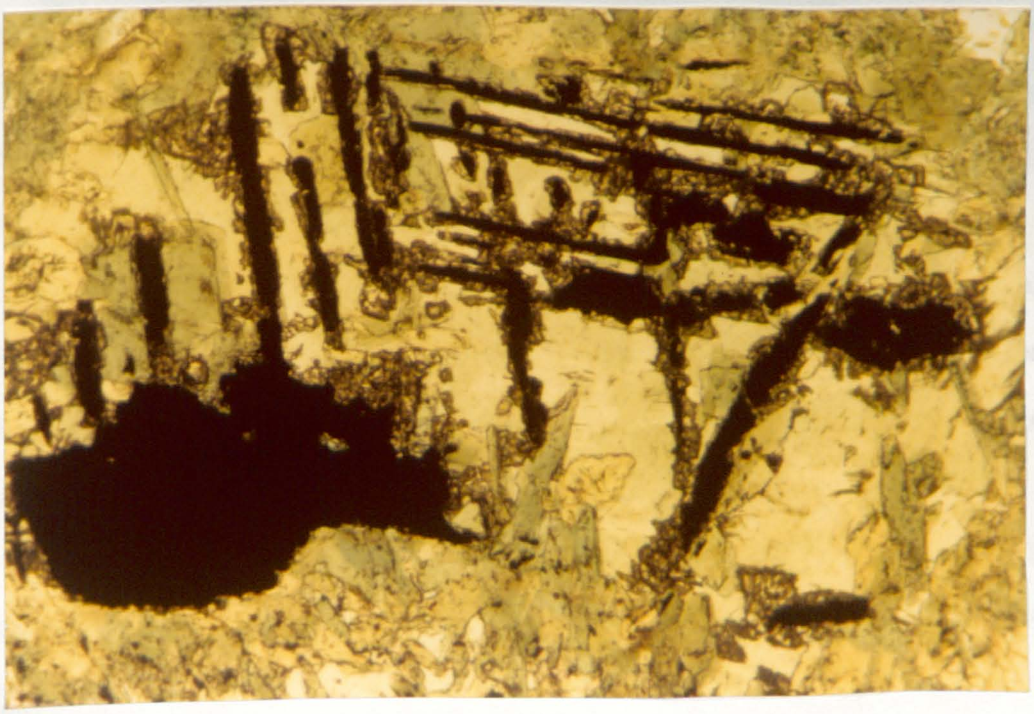


FIG. 3.12. Skeletal primary magnetite surrounded by sphene after metamorphism. PPL. X50.

to form amphibolite. As well as the large patches of sulphide (up to $1mm^2$), there are often much smaller crystals (0.1 - 0.2mm size) which have developed amphibole and biotite around them.

The sulphides were thus associated with the metamorphism and alteration of the gabbro to metagabbro. They were present early in the metamorphism of the rocks. They often surrounded the relict actinolitized pyroxenes, but were themselves often enclosed by metamorphic hornblende crystals. Hornblende reaction rims were also developed between sulphides and plagioclase, but only where the sulphide was also in contact with a pyroxene altering to amphibole. This may well be due to chemical constraints with cations such as magnesium being derived from the pyroxene.

3.3 THE AMPHIBOLITES

With the total loss of igneous texture, the metagabbros grade into the amphibolites. Two distinct groups of amphibolite correspond with Winklers (1976) division into 'high grade' andesine/labradorite (bytownite) amphibolites and 'low grade' albite/oligoclase amphibolites. The former have a relatively simple history in that they are the end products of complete metamorphism of basic rocks at mid-amphibolite facies.

The origin of the 'low grade' amphibolites is more complex. Some result from the retrogression of 'high grade' amphibolites (particularly on the margins of the intrusion). Others formed from the alteration of metagabbros and coronites, which were largely unaltered during the previous mid-amphibolite grade event, because of the paucity of water.

Lack of complete recrystallization under epidote-amphibolite facies conditions is indicated by the presence of plagioclase of two compositions and amphiboles of two compositions. All major minerals are xenoblastic. They develop either as clusters of small grains or as very ragged, highly poikiloblastic crystals. These amphibolites often develop as patches within the metagabbros, representing zones of increased water content.

The 'high grade' amphibolites contain labradorite or bytownite with hornblende (of one composition only) and more sphene but less epidote than the lower grade variants. The crystal shapes, though usually xenoblastic, are much less serrated than in the 'low grade' amphibolites and are usually equi-dimensional with triple point junctions. These 'high grade' amphibolites are more restricted in their occurrence, being confined to the margins of the intrusions and occasional shear zones within the gabbro. All the amphibolites occur in patches throughout the gabbro but are only sporadically developed on the margins and may or may not show preferred orientations of their minerals.

3.3.1 Plagioclase

Within the 'low grade' amphibolites, various generations of plagioclase may be present, Large relict laths (up to 9mm long) of igneous origin may still be discernible. They are highly sericitized and have very ragged margins. Often zoning is extreme (An_{65-30}) but it is not clear whether this is a primary magmatic feature or due to the amphibolitization. Other plagioclase (commonly albite or oligoclase) is metamorphic in origin and occurs as much shorter laths (0.5 - 0.8mm long). Where the amphibolite was derived from the retrogression of a 'high grade' amphibolite, then stubby laths of andesine (An_{35-45}) co-exist with oligoclase compositions (An_{10-25}). All plagioclases in 'low grade' amphibolites are associated with epidote and chlorite.

3.3.2 Amphiboles

The chemistry and habits of the amphiboles show much of the history of the amphibolites. Hornblendes in the 'high grade' amphibolites are green in colour while those in the 'low grade' amphibolites are blue-green. Where incomplete epidote-amphibolite grade overprint has occurred, blue-green hornblende develops as rims around green hornblende cores.

It should be noted here that the formation of actinolite cores with hornblende rims (often found in the deeper parts of the intrusion and dominantly in the coronites with minor amphibolitization) is due to a different mechanism linked to the lack of available water at these depths during metamorphism. Several workers (Miyashiro 1973, Klein 1969, Cooper and Lovering 1970) consider this rim-core relationship to be a miscibility gap between aluminium poor and aluminium rich amphiboles. Grapes and Graham (1978) consider that there is no miscibility gap and that the rims are due to prograde metamorphism. At Grøsli, it is considered that the feature is a result of differing micro-chemical environments affecting ionic diffusion. The partial pressure of water (pp H_2O) decreased inwards. In the core of the intrusion there was insufficient water to allow the necessary mobility and access of elements such as Al to form hornblende from the pyroxenes. Actinolite, which is chemically more like the pyroxenes, thus developed, except at the edges of crystals in contact with plagioclases. Here, the grain boundary permitted the movement of water so that Ca and Al from the plagioclase and Mg and Fe from the pyroxene could combine to form a hornblende reaction rim. Towards the periphery of the intrusion, there was enough waterpresent to allow the unhindered formation of hornblende.

3.3.3 Epidote

The epidotes form accessory or major components depending on the type of amphibolite. Where abundant, they form idioblastic or sub-idioblastic crystals reaching 0.6mm but usually 0-1 - 0.2mm in length. Clinozoisite and pistacite are present in equal proportions. Zoisite is rare or absent. They often occur as saussuritization products of plagioclases and, in 'low grade' amphibolites, crystals are intergrown with xenoblastic crystals of hornblende, chlorite and albite/oligoclase.

3.3.4 Other minerals

In amphibolites, biotite forms large poikilobastic plates often showing strain and containing zircons with pleochroic haloes. In the 'high grade' amphibolites, it occurs intergrown with green hornblendes. Retrogressive effects include rims of blue-green hornblende, exsolution of magnetite lamellae and conversion to chlorite. Chlorite also occurs as xenoblastic mats, often developed around amphiboles. Accessories include sphene (which may occur in greater amounts in 'high grade' amphibolites), magnetite, rutile and apatite.

Actinolite-chlorite-biotite schists occur rarely on the margins of the bodies and represent narrow shear zones. Sub-idioblastic garnet occurs rarely in all amphibolite types.

3.3.5 Sulphides

The sulphide ores associated with the amphibolites are sometimes accompanied by gangue minerals, such as chlorite, quartz, biotite and calcite and enclose all silicate minerals including the products of the epidote-amphibolite facies overprint. They were therefore introduced later than the ore found deeper in the gabbro and are related to fluid activity during metamorphism. It is notable that pyrrhotite is the overwhelmingly dominant sulphide in the coronites while pyrite-pyrrhotite assemblages occur in the amphibolites. This may well be due to the fact that the pyrite was incongruently melted in the coronites while in the amphibolites later hydrous mobilization occurred. The ore, where disseminated through amphibolites, is interstitial and often shows replacive textures and veining. The ores and gangue material are themselves cut by very late, thin veinlets of iron-rich chlorite and sericite which are found throughout the basic intrusion.

3.4 THE LATE AMPHIBOLITES

These are small intrusions in the form of thin, fine-grained dykes usually occurring in pairs, or threes, and cross-cutting all gabbro and amphibolite types. They are 'low grade' non-foliate rocks with an assemblage of blue-green hornblende-plagioclase-epidote-chlorite-quartz-biotite. The rock is however still earlier in age than the sulphides re-mobilized by fluid activity under Greenschist facies conditions mentioned in Section 3.3.5.

3.5 THE GNEISSES

The supracrustal gneisses in the Grøsli area have quartz-biotite-plagioclase (albite/oligoclase) assemblages and show continuous modal variations. Garnet, sphene, potassium feldspar, microcline perthite and muscovite are common accessories which may form up to 10% of the rock. Epidote and chlorite are common retrograde minerals and may become major constituents. Sericitic alteration of the feldspars is also common. Quartz and feldspar are dominantly granoblastic (0.1 - 0.3mm across) and often form bands of differing grain sizes, but with all exhibiting triple junctions. Biotite is brown (commonly 1 - 1.5mm long) and produces a distinct foliation. These gneisses have undergone a blasto-cataclastic recrystallization (after weak cataclastic effects) in the Grøsli area. Static biotite has occasionally grown across the groundmass. Larger grain sizes (1 - 1.5mm) occur in rocks which have not been cataclased and the quartz and feldspar crystals have irregular junctions, whilst straining is much more evident. Garnets (of the pyralspite group, Table 3.4), where present, have developed typical static growth, poikiloblastic textures.

An epidote-amphibolite facies overprint in the gneisses has produced epidote, chlorite and green biotite. Static and stress-orientated growth, as well as mimetic growth over foliations, has occurred. The biotite forms either individual laths or clusters of equi-dimensional crystals associated with epidote: it is often altered to chlorite. Rare sub-oval, relict augen of quartz and feldspar (up to 2.5mm diameter) are preserved while quartz veins are common. Late fluid alteration of the gneisses has caused heavy sericitization of the feldspars and chloritization of the biotites. Associated quartz and quartz-epidote veins contain hydrated iron oxides and or calcite.

TABLE 3.4

ELECTRON MICROPROBE MINERAL ANALYSES

	1	2	3
SiO_2	37.29	27.14	26.98
TiO_2	0.00	0.06	0.15
Al_2O_3	21.69	17.11	18.66
MgO	7.57	2.70	9.84
FeO	30.06	43.01	32.51
MnO	0.74	0.62	0.71
CaO	2.21	0.00	0.00
Na_2O	0.00	0.09	0.14
K_2O	0.00	0.05	0.05
Cr_2O_3	0.00	0.14	0.00
ZnO	N.D.	N.D.	N.D.
TOTAL	99.53	90.92	89.04

FeO = total iron as FeO

N.D. = not determined

Analysis 1 — Garnet in supracrustal gneiss. An average of five analyses.

Analysis 2 and 3 — Late stage chlorite adjacent to sulphides in altered (halo) gneiss (see Fig. 3.15).

3.5.1 Amphibolite bands in the gneisses

These thin bands (reaching a maximum width of 10cm) are a part of the supracrustal sequence. Their mineralogy is identical to that of the amphibolites formed from the basic intrusives and they contain blue-green hornblende (often poikiloblastic), abundant epidote, biotite and zoned plagioclase (An_{42-27}) with accessory quartz, sphene, magnetite and garnet.

3.5.2 Disseminated accessory sulphides

Massive sulphide ore does occur in the gneisses (see Section 3.6) but disseminated sulphides are not common. Where they are present, host rock alteration may, or may not, be present and the sulphides always show apparently replacive textures and enclose all silicates.

3.6 THE HYDROTHERMALLY ALTERED ROCKS (GNEISSES AND GABBROS)

Late solutions which effected minor redistribution of the ore have been so active in restricted parts of the mine area that total alteration of both basic and acidic host rocks has occurred. Borehole 4 cuts through a large body of massive ore, and alteration is present in a zone of sulphide dissemination for five metres on either side. Through this distance, the gneiss host rock has been progressively altered by the fluids. The initial effects were sericitization of feldspars and chloritization of micas (Fig. 3.13). Nearer the ore body, sulphide disseminations become heavier and sulphide-chlorite-muscovite aggregates are intergrown. At this stage the feldspars are totally replaced by gangue and relict quartz, zircons and garnets

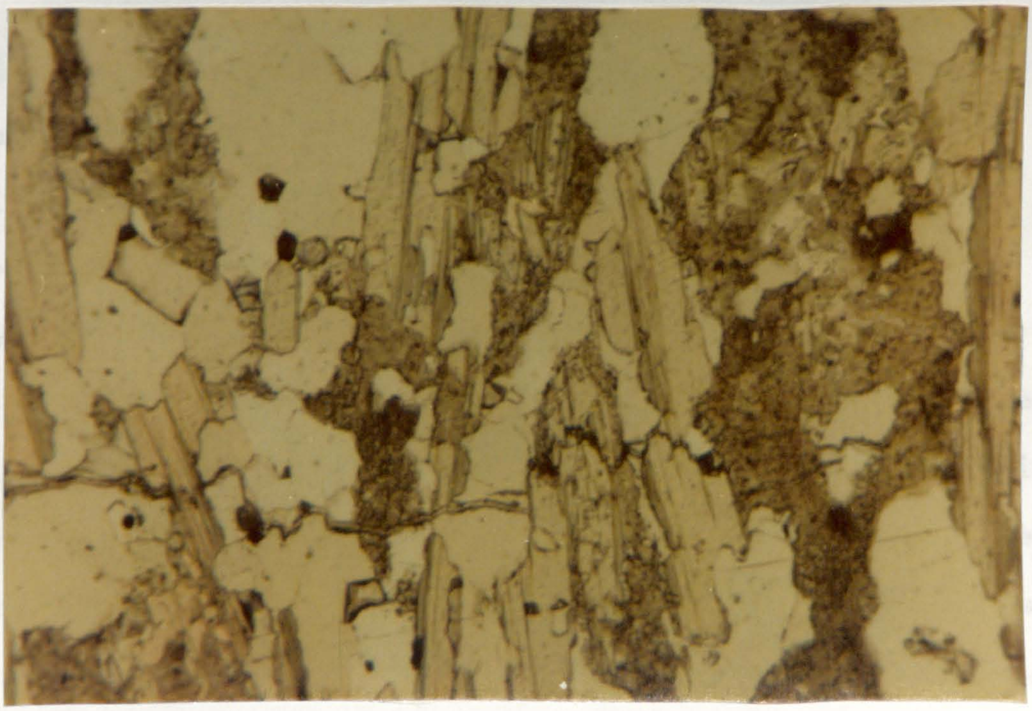


FIG. 3.13. Initial effects of alteration of gneiss. Chloritization of micas (patches) and sericitization of feldspars (laths). Unaltered material is quartz. PPL. X12.5.

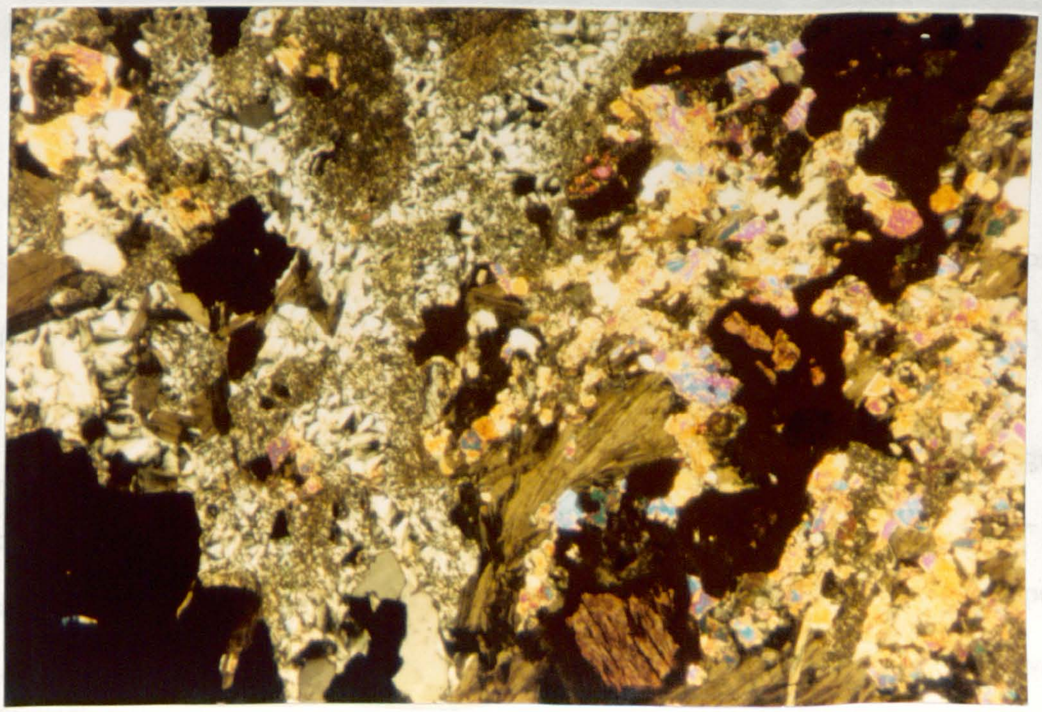


FIG. 3.14. Extensive alteration effects of gneiss, producing a sulphide-chlorite-muscovite assemblage with relict quartz crystals. XPL. X12.5.

are all that remain of the original gneiss (Fig. 3.14). The rock eventually becomes a sulphide-chlorite-muscovite/sericite-quartz-(calcite) assemblage. A later stage alteration event has formed an iron-rich chlorite-sericite assemblage (Table 3.4). Sulphides have reacted with the fluids developing embayments and ragged margins (Fig. 3.15) and veins of this late stage often cut the ore. Calcite, chloritoid and some re-mobilized chalcopyrite are also associated with this stage. Borehole 11 is the only locality in the mine area where alteration has occurred over an extensive area. Both gneiss and basic rock have been affected and have been totally replaced by an aggregate of sulphide and micas (Fig. 3.16). In places, the replacement was sufficiently static to allow the original sub-ophitic textures to be preserved: the plagioclase laths were replaced by an aggregate of muscovite, whilst the mafic minerals were replaced by a sulphide-chlorite assemblage (Fig. 3.16).

3.7 SUMMARY OF THE METAMORPHISM OF THE GABBROIC INTRUSION

During the Sveconorwegian orogeny the 'Vinor' gabbroic intrusion at Grøsli underwent mid-amphibolite facies grade metamorphism with a later overprint of epidote-amphibolite grade. The differing rock types now present (coronite, metagabbro and amphibolite) result largely from the variation in the partial pressure of water during the metamorphorphic event and the presence of fluid is necessary as a transport medium for ion exchange between minerals. The coronites are preserved only at the cores of the intrusions where water did not penetrate to any great extent. With increasing ppH_2O amphibolitization occurred.

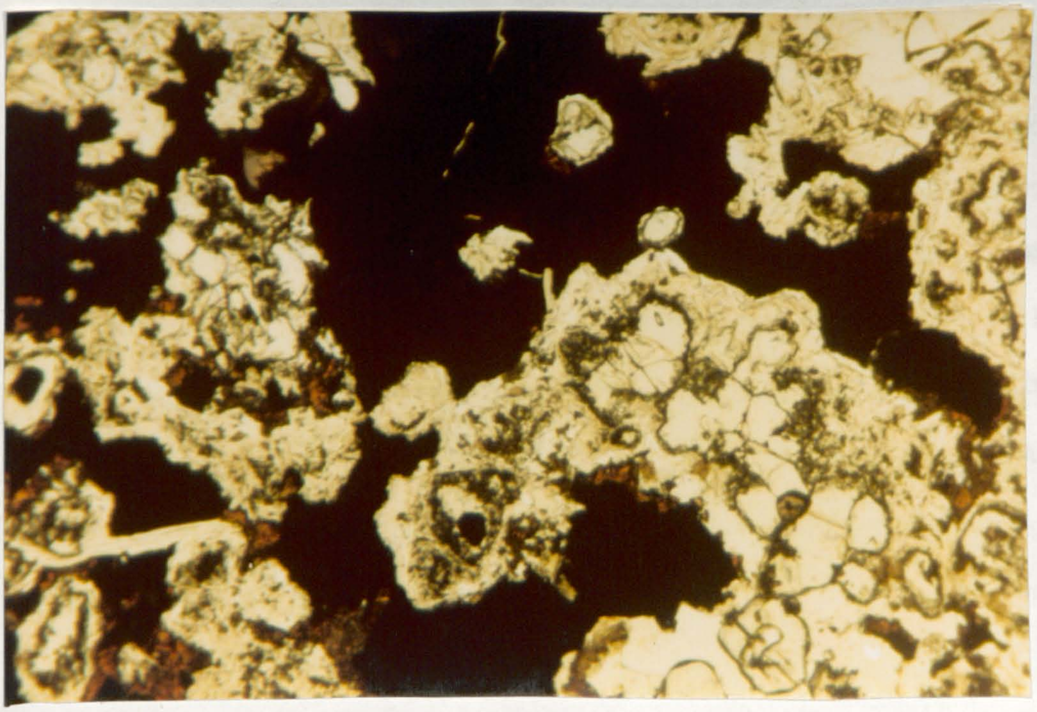


FIG. 3.15. Embayments in sulphide containing high iron chlorite and garnet. PPL. X31.5.

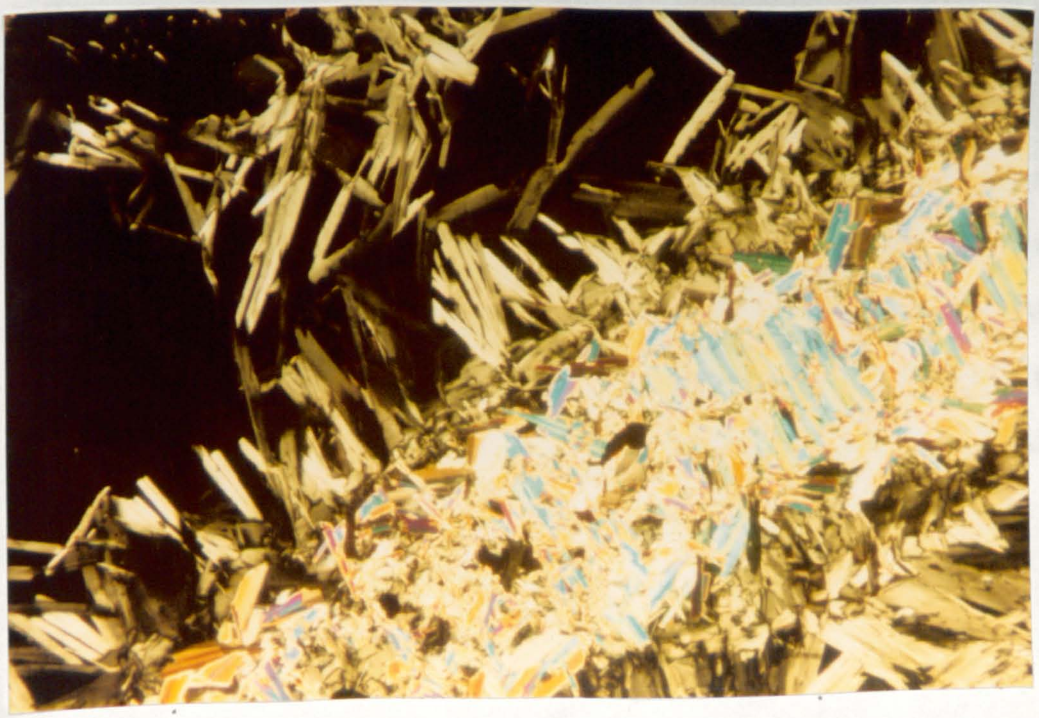


FIG. 3.16. Total replacement of coronite by sulphide-chlorite-muscovite assemblage. Muscovite crystals are pseudomorphing an original plagioclase lath. XPL. X12.5.

In many of the innermost rocks, the only alteration is the actinolitization of pyroxenes. Actinolite is not present in any of the rocks which have completely recrystallized at mid-amphibolite or epidote-amphibolite facies grade. The actinolites always have a rim of hornblende. This association is a function of chemistry under conditions of restricted water supply. The pyroxene recrystallized to chemically similar actinolite but at its margin, water present along grain boundaries permitted a reaction rim of hornblende to form against the adjacent plagioclases. This feature is considered to be formed during the mid-amphibolite facies metamorphism.

Many of the rocks show only the epidote-amphibolite facies mineralogy of epidote-albite/oligoclase-hornblende-chlorite. These assemblages are often developed over relict igneous textures with no indication of the higher calcic plagioclase-hornblende assemblage of the mid-amphibolite facies grade. This is due to the fact that only with the lower grade event did water penetrate sufficiently to cause substantial recrystallization. The amphibolites which developed at the exterior of the body were recrystallized into true mid-amphibolite grade facies metabasites since water was present in larger quantities at the periphery. The epidote-amphibolite overprint produced two kinds of rock: recrystallized epidote-amphibolites and retrogressed mid-amphibolite grade rocks still showing relict higher grade mineralogy.

Sulphides within the gabbro appear to have two origins. In some cases, the ore has interstitial relationships with the coronite minerals and there is no alteration of the host rock. It is suggested that the ore has been introduced by assimilation (i.e. engulfed by the intruding gabbro, melted and recrystallized). In the coronites, many of the sulphides also have a hornblende rim indicating the presence of the ore prior to the mid-amphibolite facies metamorphism. Some of the ore present in the metabasites was redistributed by later solutions which have variously altered the host rocks, with minor effects in the coronites varying to total replacement in the metagabbros, amphibolites and gneisses especially towards the margin of the intrusion. The coronites were thus the least affected by the re-mobilized ore and its fluids.

It is envisaged that only a small volume of ore was re-mobilized as a result of fluids passing around and through the ore bodies after the epidote-amphibolite overprint. This would imply that although the ore bodies at the periphery of the intrusion are enclosed within gangue minerals, the ore was present before the gangue minerals formed. Extensive gangue mineralization is not found away from the ore bodies and there does not seem to be any channel along which the hydrothermal fluids can be said to have passed.

In conclusion, the gabbro was intruded prior to, or at the start of, the Sveconorwegian orogeny, during which it was variably metamorphosed according to the availability of water. Sulphide ore was incorporated as the rock was intruded and was re-mobilized on a small scale after the epidote-amphibolite overprint. The volume of ore mobilized at this stage decreased with increasing distance into the gabbro body. Final fluid movement produced iron-rich chlorite-sericite veins.

CHAPTER FOUR

CHEMISTRY OF THE SILICATE ROCKS AT GRØSLI

I. THE SUPRACRUSTALS

4.1 INTRODUCTION

The supracrustal rocks in the Grøsli mine area are a monotonous sequence of leucocratic, quartzo-feldspathic gneisses, with very minor amphibolite bands. They have been thoroughly metamorphosed and now exhibit a granoblastic texture with foliation developments proportional to their mica content. No primary structures remain to indicate whether these rocks were originally volcanic (rhyolitic flows, tuffs etc.) or sedimentary (quartz-rich akoses, sandstones etc.). In the absence of such structures, the chemistry of these rocks provides the only clue as to their origin. Much chemical work has been done on distinguishing igneous and sedimentary characteristics and rocks of unknown origin have been compared with known igneous and sedimentary suites based on major and trace element data. The simplest form of correlation is the comparison of two sets of data, but more complicated and refined plots of trends in specific elements are of greater use in the problem of defining origins.

4.2 MAJOR ELEMENT ANALYSES: RELEVANT PREVIOUS WORK

The starting point of any attempt to define the origin of a group of unknown rocks is sufficiently reliable data of known igneous and sedimentary suites. There have been many studies on the compositions of igneous rocks; perhaps the most detailed being those of Daly (1933) and Nockolds (1954), who published calculated averages of a wide variety of rocks. More recently, Le Maitre (1976) has produced a set of averages for the common igneous rock groups and has also noted their variability in the form of standard deviations. Published calculations of unmetamorphosed sediments are also readily available, e.g. Pettijohn (1963), Wedepohl (1969) and Rogers and McKay (1972).

Average compositions of igneous rocks of similar chemistry to the Grøsli gneisses are reproduced in Table 4.3 while sedimentary analyses are reproduced in Tables 4.4 and 4.5. Two features are evident from the values in these tables: there is an overlap between the sedimentary and igneous fields and also between individual rock types within these fields. Due to this major problem, comparison of raw data alone is very often of little use in determining the origin of metamorphic acidic gneisses.

The plotting of various chemical elements has been widely used in attempts to reveal trends within a suite of rocks, which could then be designated as either igneous or sedimentary. In most of these plots there is an overlap between various sedimentary and igneous fields due to the limited number of elements considered. As an illustration of the problem the much used diagrams of Leake (1964; originally used to distinguish origins of amphibolites) have been reproduced as Figs. 4.1 and 4.2.

Leake's Niggli c versus mg plot (Fig. 4.1) shows the data for the Grøsli gneisses together with results for sedimentary and acid to acid-intermediate igneous rocks. Obviously this type of diagram does

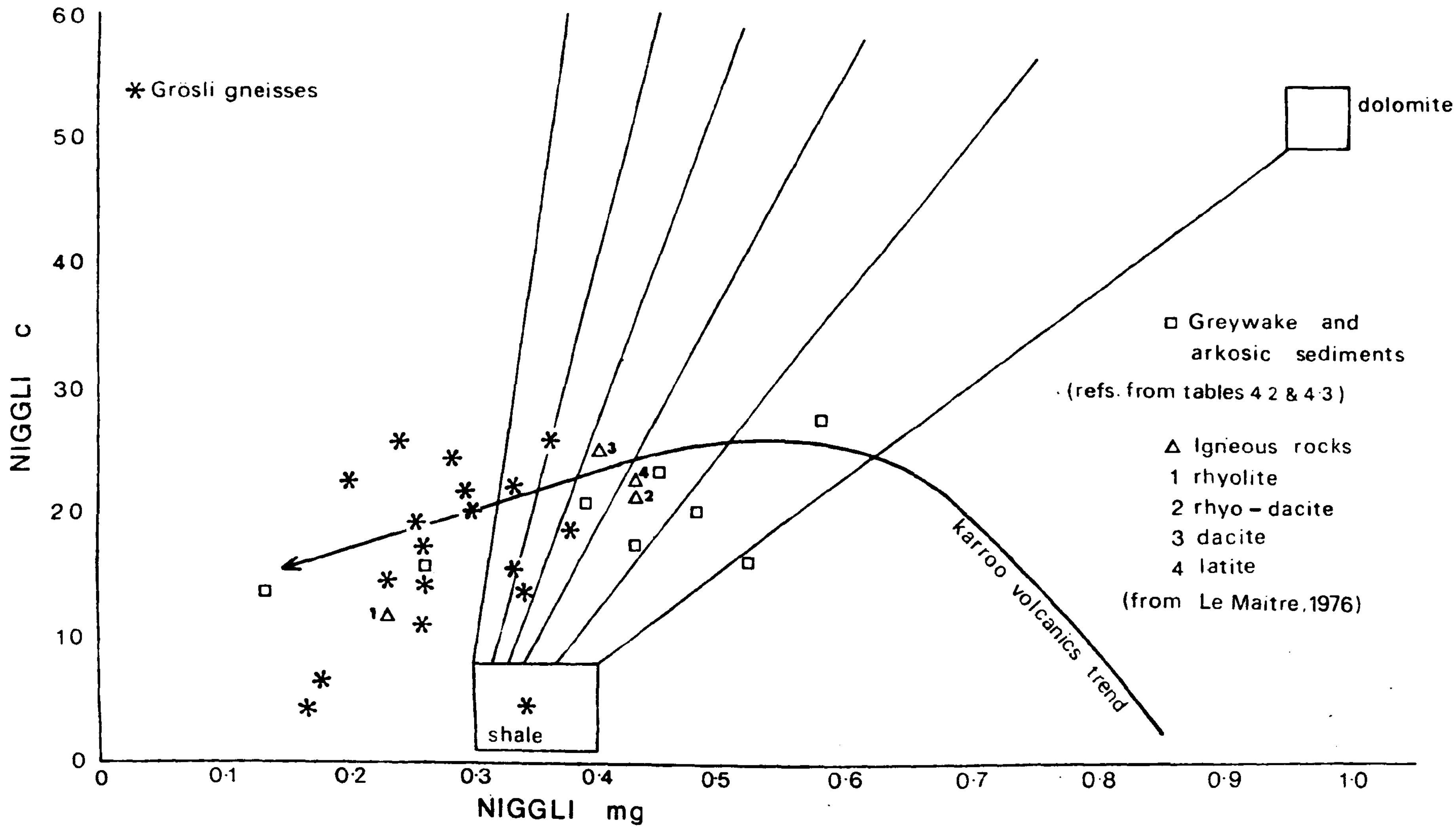


FIG. 4·1 (after Leake, 1964.)

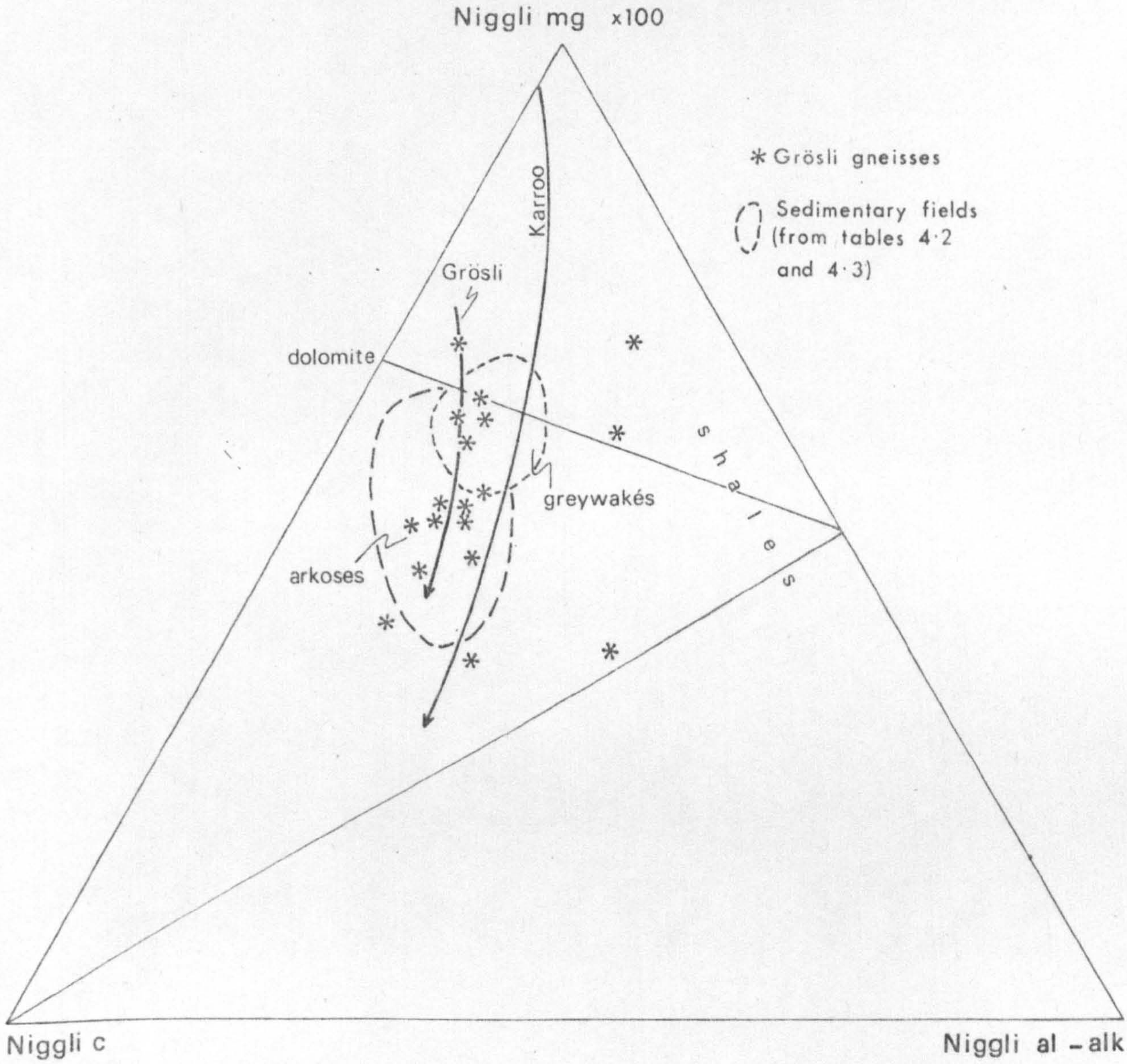


FIG. 4·2 (after Leake, 1964)

not provide sufficient discrimination. Similarly, Leake's 100mg/c/al-alk plot (Fig. 4.2) includes the Grøsli gneisses together with the sedimentary fields taken from Tables 4.4 and 4.5. The overlaps again render this diagram of little use.

Other diagrams have been extensively used to plot the fields of more acid igneous rocks and sediments which are the likely protoliths for the Grøsli gneisses. On the following diagrams the Grøsli rocks have been sub-divided as 'meta-igneous' or 'meta-sedimentary' in origin: this distinction is based on conclusions at the end of this chapter, but utilised in these plots for clarity.

A diagram of $MgO/K_2O/Na_2O$ (Fig. 4.3) produced by Sighinolfi et al (1978) also has an igneous trend line which cuts through the sedimentary fields so that any scatter casts doubts on the prediction of origins. Perhaps the best known triangular diagram for showing the differentiation trends of igneous rocks is the AFM diagram. The trends are well established for many rock suites, e.g. from the work of Nockolds and Allen (1953), Coombs and Wilkinson (1969), Carmichael (1964) and many others. In these diagrams it was generally accepted that if the analyses of the major elements of a group of rocks plotted close to known igneous trends, then the unknowns were most probably igneous in origin. The feature of iron enrichment followed by that of alkali enrichment in the complete differentiation of basaltic magmas has never been observed in sedimentary suites, and this is perhaps the best criterion for assigning igneous origins.

However, Robinson and Leake (1975) provide ample evidence that many acidic sedimentary suites (which, by their nature will plot towards the alkali apex of the AFM diagram) fall on the late stage, igneous

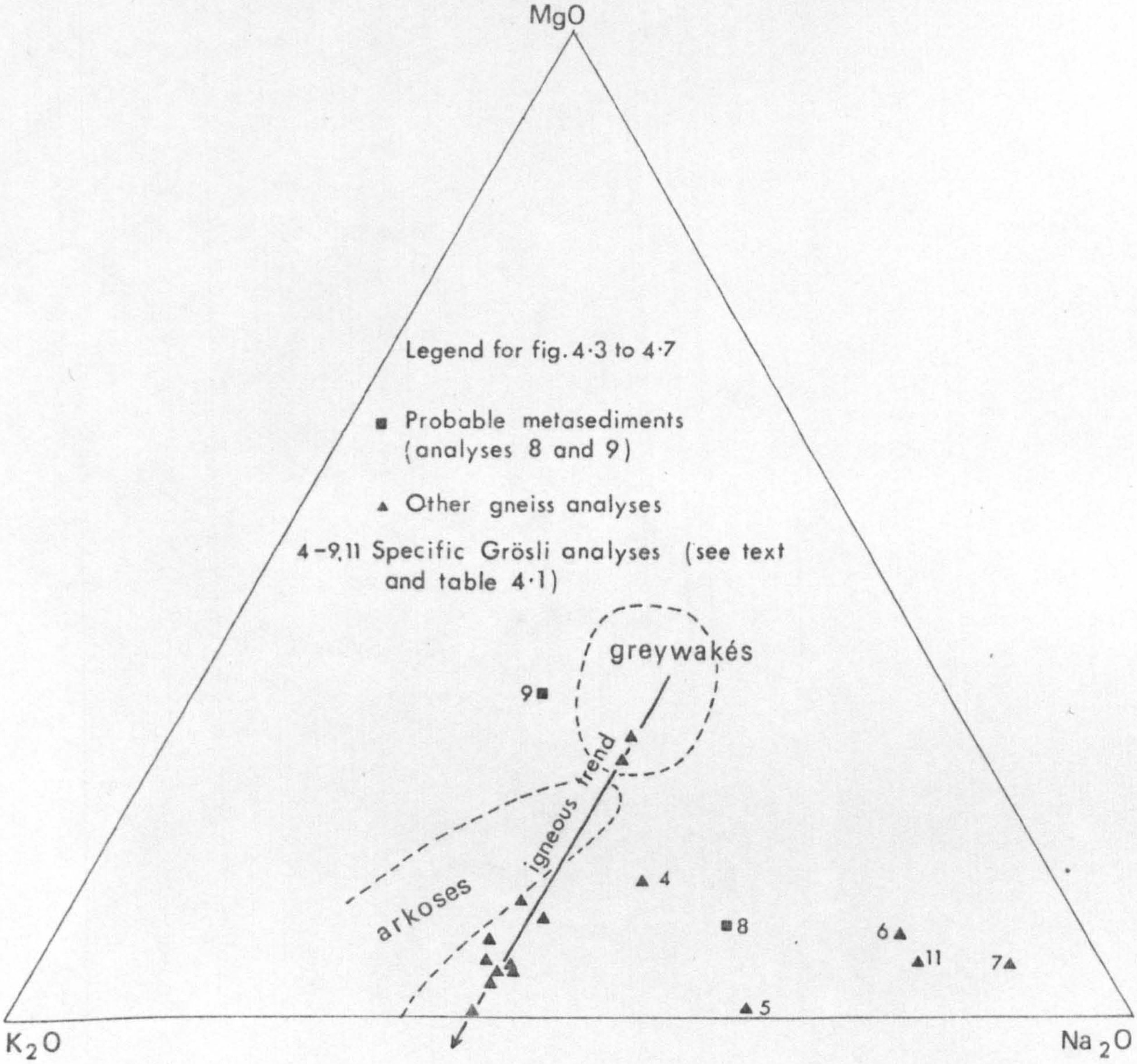


FIG. 4·3 (after Sighinolfi et al., 1978) differentiation lines. This is further complicated by the fact that calc-alkaline suites often do not show early iron enrichment trends and have no curve in their differentiation line. Such suites can also display the kind of scattering thought previously to be characteristic of sedimentary rocks. Thus, the apparent alkali enrichment of igneous rocks overlaps the fields of sandstones, arkoses and quartzo-feldspathic gneisses.

Therefore, using major element plots it is possible to suggest a sedimentary origin for points with very wide scatters or which plot in a sedimentary field away from an igneous trend line. The presence of an iron-enrichment curve is diagnostic of igneous suites, but this phenomenon is not always developed. For a large number of cases these diagrams do not provide unequivocal answers.

4.3 MAJOR ELEMENT ANALYSES OF THE GRØSLI GNEISSES

The major element chemistry of these gneisses is presented in Table 4.1. Two of the samples (nos. 1 and 2) represent hybrid rocks. They are gneisses which have been contaminated by basic material. They were formed at the junction of the gabbros and the supracrustals. Of the remaining 18 samples, 15 have silica values above 70% and all are above 60%

In a comparison of absolute values between the Grøsli rocks (Table 4.1) and analyses of igneous rocks given by Joplin (1963), all but two of the Grøsli rocks have similar chemistries to rocks in Joplin's 'granitic' groups. The compositions of these rocks range from dacites and granodiorites, through rhyo-dacites and adamellites

TABLE 4.1. MAJOR ELEMENT CHEMISTRY OF THE GRØSLI GNEISSES (weight percent)

ANALYSIS	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	TOTAL (excluding H_2O)
1	54.71	0.69	17.26	3.75	7.16	0.21	4.78	2.62	3.13	3.10	97.41
2	52.54	0.56	17.06	1.44	7.89	0.19	6.06	7.69	3.49	1.54	98.46
3	61.93	1.45	15.40	3.68	5.27	0.13	2.01	4.33	3.10	2.46	99.76
4	67.37	0.95	13.95	2.10	3.11	0.09	1.07	3.78	3.76	2.82	99.00
5	77.92	0.07	12.22	0.00	0.64	0.01	0.07	0.49	5.30	2.77	99.49
6	77.23	0.30	11.47	0.72	0.94	0.02	0.51	2.50	4.34	0.95	98.98
7	75.25	0.22	13.11	0.28	0.94	0.01	0.39	1.79	6.06	0.59	98.64
8	68.76	0.73	13.11	3.75	1.80	0.10	0.74	3.25	4.31	2.33	98.88
9	73.01	0.49	11.22	2.62	4.49	0.47	2.14	0.68	1.93	2.29	99.34
10	73.27	0.20	12.17	1.90	3.56	0.04	0.61	0.48	2.98	4.06	99.27
11	76.24	0.22	13.13	0.63	1.16	0.02	0.39	2.08	5.11	1.10	100.08
12	71.50	0.37	12.15	2.94	3.93	0.17	1.22	3.95	1.71	1.29	99.23
13	72.14	0.39	12.95	0.91	1.81	0.05	0.51	2.19	3.46	4.72	99.13
14	75.48	0.23	12.49	0.20	1.03	0.03	0.31	1.42	3.52	4.74	99.45
15	71.02	0.52	13.34	0.99	2.19	0.06	0.84	2.76	3.41	3.83	98.96
16	74.55	0.20	13.00	0.32	1.42	0.04	0.36	1.38	3.50	4.33	99.10
17	75.40	0.25	13.53	0.96	1.31	0.03	0.38	1.52	3.63	4.74	101.75
18	74.26	0.31	12.63	0.87	1.37	0.03	0.41	1.73	3.26	4.12	98.99
19	74.78	0.21	12.54	0.53	0.97	0.03	0.03	1.02	3.47	4.95	98.53
20	71.36	0.56	13.36	1.32	2.38	0.07	1.05	2.26	3.41	4.23	100.00

TABLE 4.1 continued

1 and 2.	Hybrid rocks. Basified gneisses (see text).
8 and 9.	Probable para-gneisses.*
3 to 7 and 10 to 20.	Probable ortho-gneisses.*

*Distinctions based on conclusions of Chapter Four.

TABLE 4.2. TRACE ELEMENT CHEMISTRY OF THE GRØSLI GNEISSES (ppm)

ANALYSIS	Cr	V	CO	Ni	Cu	Zn	Cd	Rb	Sr	Y	Zr	Ba	Ce
1						NOT DETERMINED							
2						NOT DETERMINED							
3	66	78	14	24	21	125	62	85	242	64	410	845	76
4	<50	86	15	6	16	96	3	78	239	36	182	552	38
5	<50	5	N.D.	2	3	16	N.D.	50	74	23	59	266	11
6	<50	17	4	9	16	23	N.D.	7	154	20	205	215	23
7	<50	16	2	5	4	53	N.D.	7	183	15	346	156	60
8	<50	31	16	7	85	77	N.D.	46	185	51	119	628	61
9	<50	63	10	36	478	806	N.D.	36	19	21	71	255	12
10	<50	26	4	8	53	166	5	101	63	24	115	479	47
11	<50	17	4	18	12	27	N.D.	23	209	26	127	248	58
12	<50	6	7	7	15	224	N.D.	26	91	17	114	202	16
13	99	24	7	8	8	39	N.D.	105	152	42	227	601	72
14	89	16	4	5	4	53	N.D.	131	109	28	144	523	39
15	<50	40	9	8	1	45	N.D.	110	164	42	186	566	49
16	<50	14	2	2	1	31	N.D.	116	131	30	157	549	54
17	<50	18	3	4	2	28	N.D.	124	120	29	141	541	57
18	94	29	6	6	8	92	N.D.	56	162	23	181	768	80
19	<50	13	2	3	1	25	1	126	103	29	240	539	22
20	<50	40	12	12	12	94	1	132	124	27	166	479	28

N.D. Not detected

to rhyolites, quartz-keratophyres and granites. One very high silica value from Grøsli (no. 5) is similar to a felsite. Only two analyses from Grøsli (nos. 8 and 9) apparently have no close comparison with the igneous rocks of Joplin. Analysis 2 (of gneiss contaminated by gabbro) is very similar to a quoted analysis of diorite which had been formed by the introduction of basic material into a pelite.

A comparison of the Grøsli analyses (Table 4.1) with the averaged analyses of Le Maitre (1976 and Table 4.3) indicates that most of the Grøsli gneisses fall within the rhyolite field, with a few less acidic members tending towards the rhyo-dacite field. One sample (no. 3) has a low enough silica content to appear in the latite composition range. Analyses 8 and 9 again have no similarities with any of the igneous groups.

As previously noted, the compositional fields of igneous and sedimentary rocks have considerable overlaps. The Grøsli gneisses can be also compared with the sandstone-arkose-greywacke groups of sediments (Tables 4.4 and 4.5). An upper limit of 69% SiO_2 (Andreae, 1974) for the greywacke group precludes all but three analyses from Grøsli (nos. 3, 4 and 8). Analysis 3 has rather a high TiO_2 content while analysis 8 has a rather low magnesium value. Analysis 4 has the most similar composition to the group. The remaining rocks (i.e. nos. 1, 2, 5, 6, 7, 9 to 20) all fall within the arkose-sandstone fields. Since analyses 8 and 9 correlate with the wider ranges of the sedimentary fields, whereas they do not fit into the igneous fields, it seems reasonable to propose a sedimentary origin for these rocks. However for the other rocks on the basis of these comparisons alone, a sedimentary or an igneous origin could be assigned.

TABLE 4.3

AVERAGE IGNEOUS ROCK COMPOSITIONS

ANALYSIS ($\pm$ 2σ)	1	2	3	4
SiO_2	74.00 $\pm$ 3.51	67.52 $\pm$ 4.61	66.36 $\pm$ 4.17	62.80 $\pm$ 5.63
TiO_2	0.27 $\pm$ 0.25	0.60 $\pm$ 0.30	0.58 $\pm$ 0.32	0.83 $\pm$ 0.40
Al_2O_3	13.53 $\pm$ 1.77	15.53 $\pm$ 1.36	16.12 $\pm$ 1.67	16.37 $\pm$ 1.45
Fe_2O_3	1.47 $\pm$ 1.24	2.46 $\pm$ 1.35	2.39 $\pm$ 1.31	3.34 $\pm$ 1.62
FeO	1.16 $\pm$ 1.29	1.80 $\pm$ 1.20	2.41 $\pm$ 1.55	2.27 $\pm$ 1.54
MgO	0.41 $\pm$ 0.48	1.68 $\pm$ 1.33	1.74 $\pm$ 1.14	2.25 $\pm$ 1.86
CaO	1.16 $\pm$ 0.96	3.35 $\pm$ 1.53	4.29 $\pm$ 1.65	4.27 $\pm$ 1.79
Na_2O	3.62 $\pm$ 1.29	3.90 $\pm$ 0.70	3.89 $\pm$ 0.91	3.88 $\pm$ 0.77
K_2O	4.38 $\pm$ 1.69	3.16 $\pm$ 0.78	2.22 $\pm$ 1.00	3.98 $\pm$ 0.90

1. Rhyolite. Average of 116 analyses.
2. Rhyo-dacite. Average of 40 analyses.
3. Dacite. Average of 80 analyses.
4. Latite. Average of 46 analyses.

All from Le Maitre (1976).

TABLE 4.4

AVERAGE ARKOSE COMPOSITIONS

ANALYSIS ($\pm$ 2σ)	1	2	3	4
SiO_2	76.37 $\pm$ 5.12	69.62 $\pm$ 3.59	71.86 $\pm$ 5.53	69.56 $\pm$ 2.05
TiO_2	0.41 $\pm$ 0.01	0.44 $\pm$ 0.11	0.37 $\pm$ 0.13	0.59 $\pm$ 1.01
Al_2O_3	10.63 $\pm$ 2.09	12.84 $\pm$ 1.57	12.34 $\pm$ 3.18	13.44 $\pm$ 1.02
Fe_2O_3	2.12 $\pm$ 1.30	1.78 $\pm$ 0.92	1.55 $\pm$ 0.66	5.01 $\pm$ 1.12
FeO	1.22 $\pm$ 0.46	1.65 $\pm$ 1.03	1.10 $\pm$ 0.37	As Fe_2O_3
MnO	0.38 $\pm$ 0.46	0.06 $\pm$ 0.03	0.05 $\pm$ 0.01	0.06 $\pm$ 0.02
MgO	0.30 $\pm$ 0.29	1.73 $\pm$ 0.75	1.16 $\pm$ 0.33	0.89 $\pm$ 0.34
CaO	1.30 $\pm$ 1.26	2.67 $\pm$ 2.03	3.01 $\pm$ 0.60	2.01 $\pm$ 0.60
Na_2O	1.84 $\pm$ 1.23	3.07 $\pm$ 0.61	2.79 $\pm$ 0.68	2.45 $\pm$ 0.49
K_2O	4.99 $\pm$ 1.05	2.75 $\pm$ 0.78	2.80 $\pm$ 0.83	4.80 $\pm$ 0.75

1. Average of 5 arkoses from Pettijohn (1954).
2. Average of 26 arkoses from Van de Kamp et al (1976).
3. Average of 6 Holocene sands from Van de Kamp et al (1976).
4. Average of K rich metasediments from South Norway. From Andreae (1974).

TABLE 4.5

AVERAGE GREYWACKE COMPOSITIONS

ANALYSIS ($\pm$ 2σ)	1	2	3	4
SiO_2	62.25 $\pm$ 5.10	61.86 $\pm$ 1.73	63.09 $\pm$ 2.16	67.34 $\pm$ 1.44
TiO_2	0.51 $\pm$ 0.27	0.40 $\pm$ 0.04	0.58 $\pm$ 0.06	0.72 $\pm$ 0.27
Al_2O_3	14.24 $\pm$ 2.72	14.61 $\pm$ 1.69	13.91 $\pm$ 2.15	14.01 $\pm$ 1.65
Fe_2O_3	1.55 $\pm$ 1.20	5.08 $\pm$ 0.61	6.53 $\pm$ 1.21	5.85 $\pm$ 2.48
FeO	3.66 $\pm$ 1.66	As Fe_2O_3	As Fe_2O_3	As Fe_2O_3
MnO	0.14 $\pm$ 0.08			0.06 $\pm$ 0.04
MgO	2.18 $\pm$ 0.95	3.58 $\pm$ 0.55	3.61 $\pm$ 0.61	1.84 $\pm$ 0.93
CaO	2.73 $\pm$ 1.73	5.74 $\pm$ 1.31	3.05 $\pm$ 0.87	3.25 $\pm$ 1.18
Na_2O	3.02 $\pm$ 0.81	4.10 $\pm$ 0.48	3.10 $\pm$ 0.76	3.06 $\pm$ 0.68
K_2O	1.84 $\pm$ 0.42	1.90 $\pm$ 0.47	2.08 $\pm$ 0.41	1.97 $\pm$ 0.74

TABLE 4.5 continued

1. Average of 23 greywackes from Pettijohn (1957).
2. Average of 10 Precambrian greywackes from Naqvi et al (1972).
3. Average of 7 Precambrian greywackes from Naqvi et al (1972).
4. Average of metagreywackes from South Norway. From Andreae (1974).

Fig. 4.3 shows a plot of Grøsli analyses on the $MgO/Na_2O/K_2O$ of Sighinolfi et al (1978). In this diagram the majority of the points are outside the arkose and greywacke fields, suggesting the possibility of an igneous origin. Towards the Na_2O corner are a group of seven more soda-rich acidic rocks. One of these (no. 9) appears to be sedimentary in origin (see above), while the remaining five points (nos. 4, 5, 6, 7 and 11) could have been derived from soda-rhyolites. However, the mobility of alkalies during metamorphism may also account for their apparent isolation from the other points.

Fig. 4.4 shows the Grøsli gneiss analyses plotted on an AFM diagram and indicates a strongly differentiated nature, if they are of an igneous origin. The majority of the rocks plot outside the fields of arkoses and greywackes, which are possible protoliths from comparison of silica contents alone. However, Robinson and Leake (1975) have proved that analyses of sedimentary suites can extend towards the alkali apex: the arkose and greywacke fields of Fig. 4.4 could then well be enlarged and encompass the Grøsli plots.

Only a few definite conclusions can be drawn from this set of diagrams. Analyses 8 and 9 almost certainly represent original sedimentary rocks. The other analyses certainly cluster very close to igneous trends (e.g. on Fig. 4.4) but they also fall in the overlap with sedimentary fields.

Finally, Van de Kamp et al (1976) state that a plot of Niggli mg against Niggli si for sandstones and arkoses gives a positive correlation due to the concentration of mafic material in sandstones with a high silica content, whereas the converse is true in igneous rocks. This plot should be the most applicable to the Grøsli gneisses.

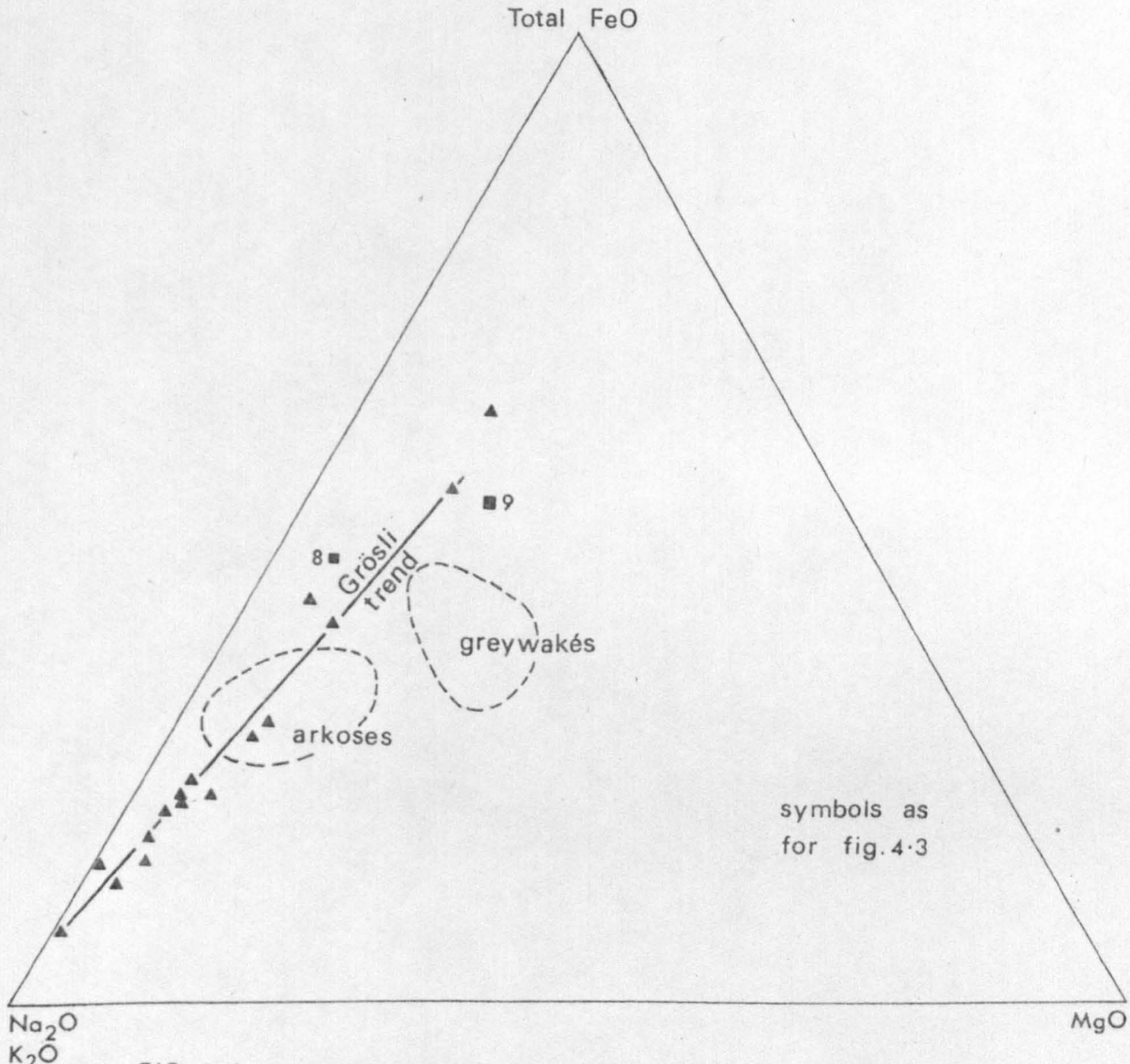

Total FeO
Grösli trend
8
9
greywakés
arkoses
symbols as for fig.4·3
Na_2O K_2O
MgO

FIG.4·4

Unfortunately the scatter of points (Fig. 4.5) is so large as to permit neither positive nor negative slopes to be drawn through the data.

4.4 TRACE ELEMENT ANALYSES: RELEVANT PREVIOUS WORK

Van de Kamp et al (1976) in studying shales, sandstones and arkoses, concluded that there was no definite trend of Ni or Cr versus Niggli mg, whereas igneous rocks displayed a more systematic variation. They stated that the behaviour of Zr is diagnostic. In igneous rocks, Zr contents rise with increasing differentiation (since Zr is an incompatible element) and then drop at the extreme acid end with the precipitation of zircon. In sedimentary rocks, the variation of Zr with si or mg is much more variable, since Zr can occur in detrital zircon in rock fragments containing zircon, or within the clay minerals. Thus a plot of Zr against si or mg will reveal a positive slope for igneous suites but no such trend for sedimentary rocks.

4.5 TRACE ELEMENT ANALYSES OF THE GRØSLI GNEISSES

Plots of the major element analyses showed that samples 8 and 9 could well have formed from sedimentary protoliths. The other specimens were probably igneous in origin, but their chemistry overlapped sedimentary fields.

Using this probable division of protoliths, a plot of Niggli Si versus Zr ppm (Fig. 4.6) shows the meta-igneous rocks broadly defining a negative slope, but with some scatter.

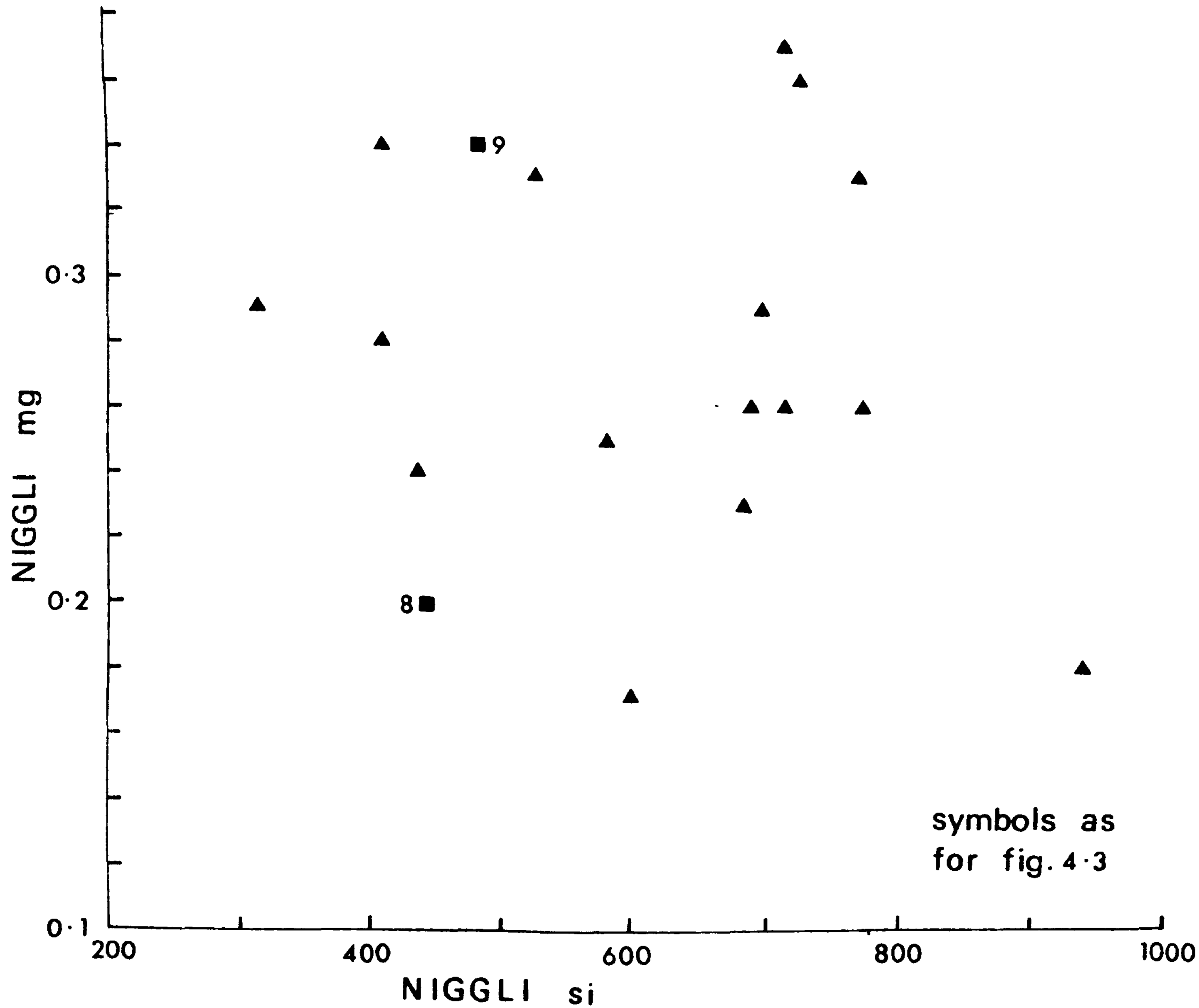

NIGGLI mg
0·3
0·2
0·1
200
400
600
800
1000
NIGGLI si
9
8
symbols as
for fig. 4·3

FIG. 4·5

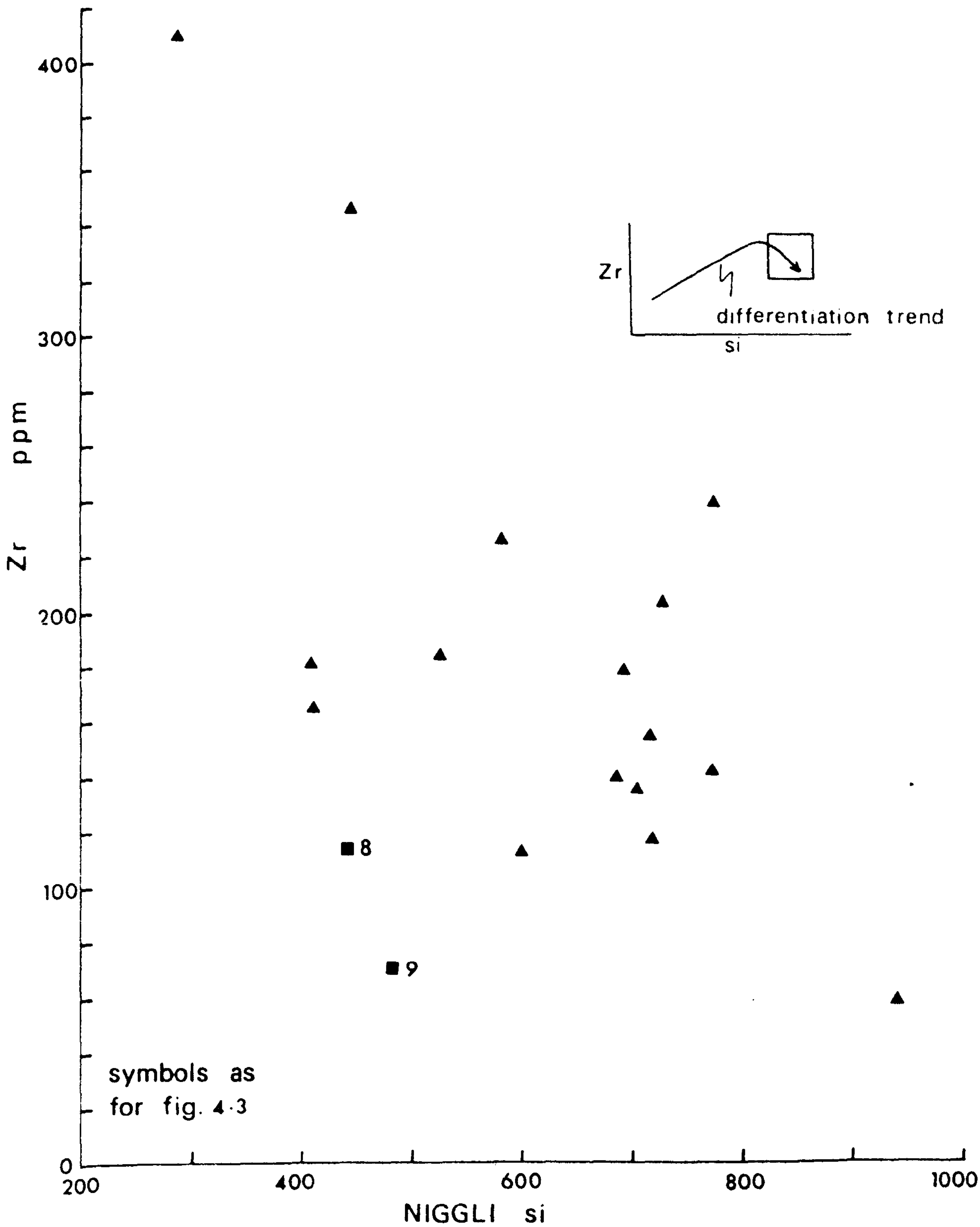

Zr ppm
NIGGLI si
400
300
200
100
0
200
400
600
800
1000
8
9
symbols as
for fig. 4·3
Zr
si
differentiation trend

FIG. 4·6

It appears that the gneisses represent highly differentiated end members in which the Zr content has started to fall; a feature suggested by Nockolds and Allen (1953) and Van de Kamp (1976).

The problem of an igneous or sedimentary origin for the gneisses has not been completely solved with the use of these diagrams. Some of the rocks show sedimentary characteristics, but the majority have igneous affinities. The igneous group may well be of two parentages (e.g. see Fig. 4.3) but metamorphism has undoubtably caused some element mobility. Under such conditions the scatter of points on the diagrams is not surprising and some of the supracrustals in the Grøsli area may now only partially retain their original chemical characteristics. However, it is concluded that the majority of the gneisses most probably represent acid volcanics while a subordinate number of quartz-rich sediments are also present.

4.6 ORIGIN OF THE GRØSLI GNEISSES

Some of the gneisses at Grøsli are suggested to have formed from acidic volcanic rocks. C. Bugge (1917) and J. Bugge (1943) considered the acidic rocks in the Kongsberg area to be of dacitic composition and probably representative of tuffs and other effusives interlayered with sediments. Starmer (1977) considered the gneisses in the south of the Kongsberg Series to be mixed metavolcanics and metasediments. In the Grøsli mine area the present study has shown many of the rocks could be metavolcanics of the rhyolite-dacite series.

Since the ore deposit at Grøsli is not of a type normally associated with basic magmatism (see Chapter Eleven), it may well have been related to the supracrustals. If this is the case, then the gneisses and ore-body may represent the acidic differentiates of a more basic magma. It would be advantageous therefore to know the environment of formation of the magma. Fox (1979) presented a discriminant function analysis of major elements in acid end members of a basic series associated with massive volcanogenic sulphides. In the analysis he separated calc-alkaline trends from those shown by tholeiitic rocks of the island-arc type. A similar diagram is produced for the Grøsli gneisses in Fig. 4.7 (which plots 'fcn.1' and 'fcn.2' of Fox, 1979).

The danger of using major element plots for metamorphic rocks is evident from the previous discussion although Fig. 4.7 places less heavy weighting on the more mobile elements such as alkalies. This fact may help to reduce variations produced by metamorphic mobility. Fig. 4.7 shows that there is some division between the supposed volcanic rocks and the supposed sedimentary representatives. On the basis of this diagram, the volcanics are mainly of calc-alkaline type, produced at destructive plate margins. This conclusion is substantiated by plots of the gneisses on Fig. 4.8 which indicates that the rocks seem to belong to a compressional environment (see Section 4.8 for description of Fig. 4.8).

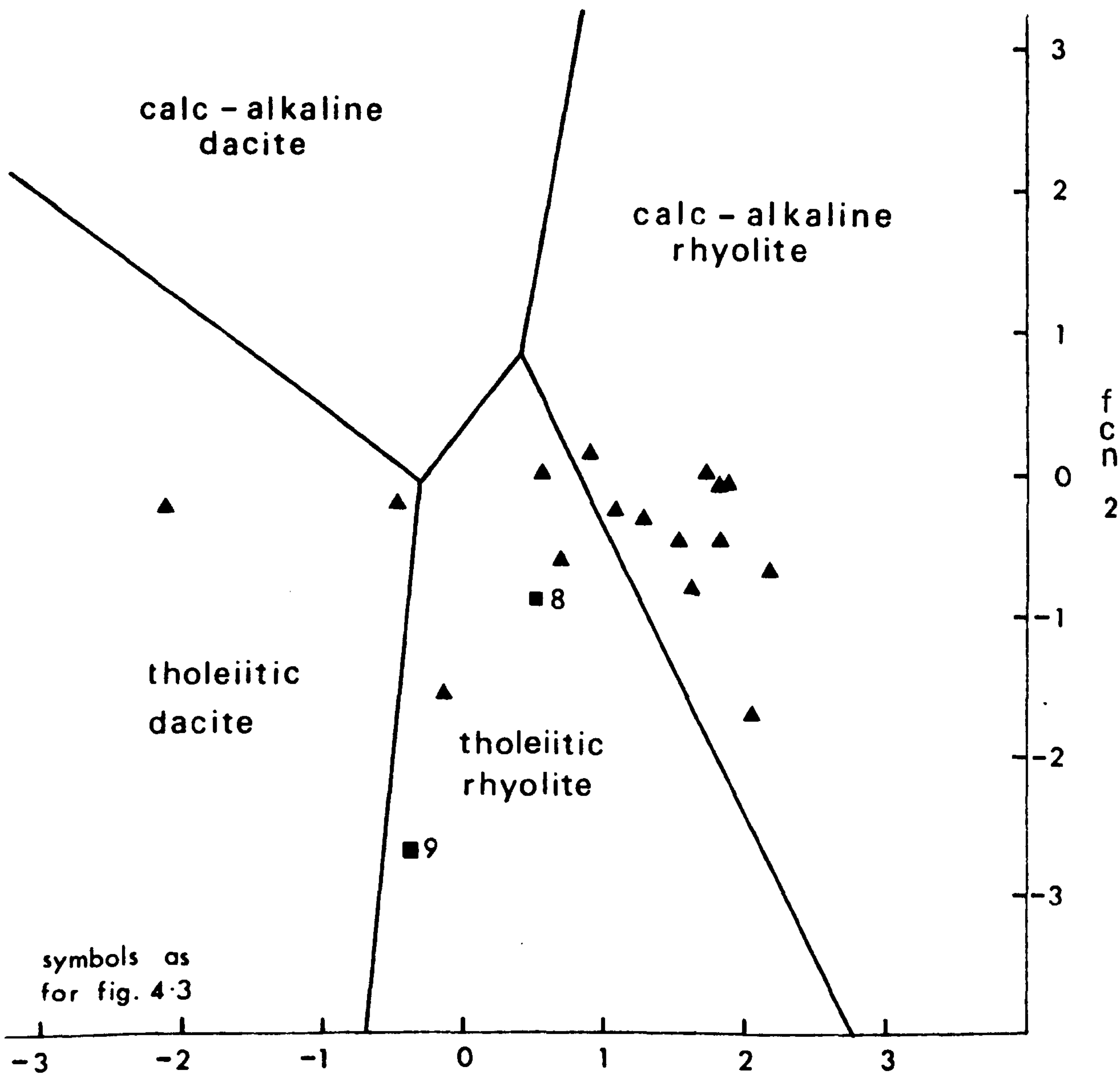


FIG. 4·7
(after Fox, 1979)

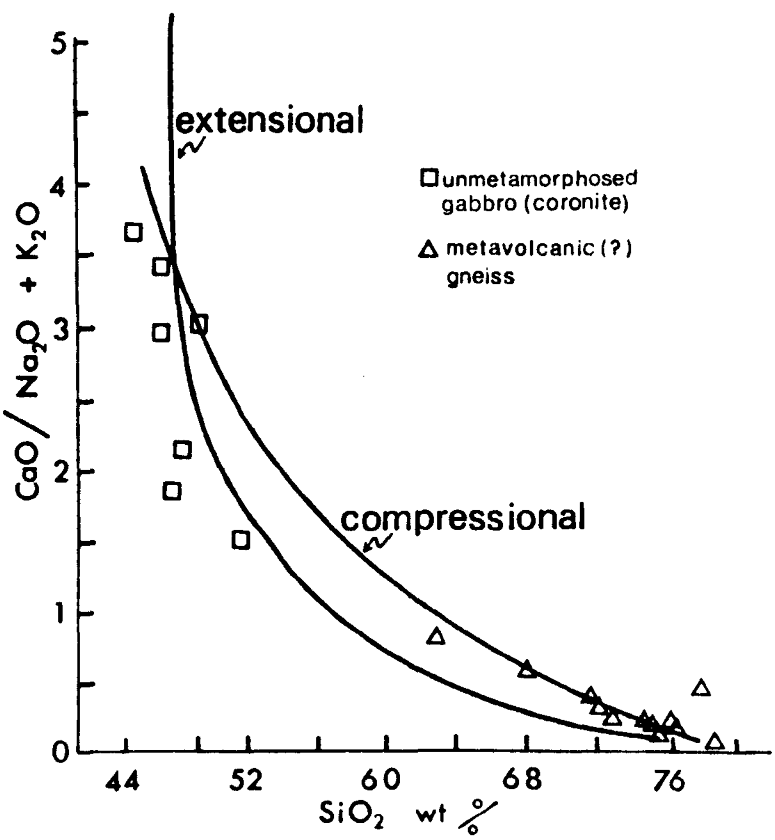


FIG. 4·8 (from Petro et al. 1979)

II. THE 'VINOR' GABBROIC ROCKS

4.7 INTRODUCTION

This group of rocks includes all the younger basic intrusives which have cut the supracrustals. Three intrusive phases are present at Grøsli (see Chapter Two); the earliest forming large bodies and the later two stages being only of low volume. There is also a good textural and mineralogical progression from coronite through metagabbro to amphibolite, in most outcrops and borehole sections. Therefore, the relationship between such rocks is usually not in question. The following element plots will attempt to assign the intrusives to one of the main basaltic groups. (An extension of the use of these diagrams was also applied in an attempt to assign the supracrustal 'volcanic' gneisses to a parent group, assuming them to be a fractionation series.) Chemical analyses of the 'Vinor' gabbros are presented in Tables 4.6 and 4.7.

4.8 MAJOR ELEMENT ANALYSES: RELEVANT PREVIOUS WORK

Analytical studies of plutonic rock suites (with relation to their tectonic setting) are not as numerous as those of volcanic suites. Three of the most recent are those by Christiansen and Lipman (1972), Johnston et al (1976) and Petro et al (1979). Work on volcanic suites has been extensive and the chemical techniques used have been applied to plutonic series. The recognition and distinction of differing suites has continued from the early work of Daly (1911) and Cross (1915).

TABLE 4.6. MAJOR ELEMENT CHEMISTRY OF THE GRØSLI 'VINOR' GABBROS (weight percent)

ANALYSIS	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	TOTAL (excluding H_2O)
1	47.90	1.63	18.29	1.38	10.43	0.17	6.87	8.86	2.68	1.51	99.72
2	47.34	1.59	17.87	1.90	9.91	0.17	7.61	8.08	2.12	2.43	99.02
3	46.94	1.61	20.00	2.06	9.43	0.17	6.39	9.60	2.69	0.57	99.46
4	45.32	1.03	19.05	1.13	9.67	0.15	11.26	8.91	2.19	0.29	99.00
5	49.02	1.52	16.95	2.68	9.34	0.20	6.63	8.58	2.41	0.44	97.77
6	47.05	1.63	18.54	1.21	11.48	0.19	7.13	9.52	2.47	0.34	99.56
7	49.43	2.31	14.14	4.32	10.07	0.25	4.50	9.72	1.76	0.85	97.35
8	48.73	1.94	15.17	2.03	11.46	0.16	6.43	8.32	3.21	0.41	97.86
9	46.03	1.30	15.57	4.74	8.72	0.18	9.21	7.33	1.83	2.77	97.68
10	46.57	1.46	16.83	2.08	8.99	0.15	8.10	8.47	2.43	2.02	97.10
11	50.52	3.13	15.65	2.12	10.34	0.20	3.43	8.27	3.21	2.40	99.27
12	46.97	1.00	20.65	1.82	8.38	0.14	7.66	9.00	2.58	0.47	98.67
13	51.70	1.20	18.32	2.81	9.12	0.36	3.61	6.66	3.07	1.97	98.82
14	49.78	2.41	13.56	3.66	10.55	0.20	5.20	8.30	2.30	1.61	97.57
15	51.28	1.57	15.54	1.71	9.23	0.17	6.70	7.03	2.47	1.53	97.23
16	47.46	1.43	17.27	1.80	9.71	0.17	7.71	7.88	2.22	2.19	97.84
17	50.18	2.75	18.08	4.99	5.63	0.14	3.12	6.16	4.62	1.57	97.24
18	52.07	1.62	17.67	1.44	9.09	0.16	5.38	7.70	2.54	1.14	98.81

TABLE 4.6 continued

1 to 6 and analysis 12.	Coronites
7 and 8.	Late stage amphibolites.
9 to 11 and 13 to 18.	Metagabbros and amphibolites.

TABLE 4.7. TRACE ELEMENT CHEMISTRY OF THE GRØSLI 'VINOR' GABBROS (ppm)

ANALYSIS	Cr	V	Co	Ni	Cu	Zn	Cd	Rb	Sr	Y	Zr	Ba	Ce
1	282	144	64	157	96	102	42	18	210	30	123	234	36
2	258	157	71	154	110	101	60	44	233	32	134	273	40
3	205	144	72	178	118	101	38	12	230	33	118	194	40
4	164	90	101	438	28	55	43	6	279	22	75	120	27
5	294	255	44	88	62	288	36	7	195	14	24	174	32
6	302	222	64	151	76	139	46	8	213	17	49	143	30
7	165	192	22	34	424	750	67	7	219	29	74	152	44
8	186	211	76	122	123	133	26	9	81	49	172	166	43
9	327	156	96	177	24	152	56	51	208	27	95	348	39
10	152	131	59	212	43	110	53	6	262	47	108	334	113
11	82	204	35	29	45	149	47	24	180	70	208	407	60
12	147	96	58	198	68	372	65	6	305	24	93	212	27
13	89	254	60	31	42	299	49	86	235	37	83	598	29
14	284	329	36	52	11	552	55	4	141	47	113	454	57
15	338	141	62	130	53	123	54	45	195	49	229	585	36
16	177	155	61	127	90	275	66	5	303	59	144	573	29
17	84	174	84	131	766	275	57	68	334	169	771	214	136
18	150	143	36	74	47	125	45	27	302	45	208	455	49

Genetic implications of chemical variation trends have slowly been recognized, with workers such as Macdonald (1960, 1964) distinguishing continental and oceanic characteristics and Kuno (1960) recognizing high alumina basalts, tholeiitic basalts and alkali basalts. The recognition of differences between 'orogenic' types (island-arc and other subduction zone volcanics) and 'anorogenic' types (tensional plate margin and Hawaii-type, within-plate magmatics) based on major element trends have been discussed by Irvine and Baragar (1971), Miyashiro and Shido (1975), Pearce (1976) and various other authors.

Petro et al (1979) combined the methods of several workers and produced a set of plots which they used in combination, to distinguish compressional and extensional environments of origin. They used plots of alkalies versus silica (from Christiansen et al, 1972), differentiation indices (from Thornton et al, 1960) and AFM plots distinguishing tholeiitic and calc-alkaline series (from Irving and Baragar, 1971).

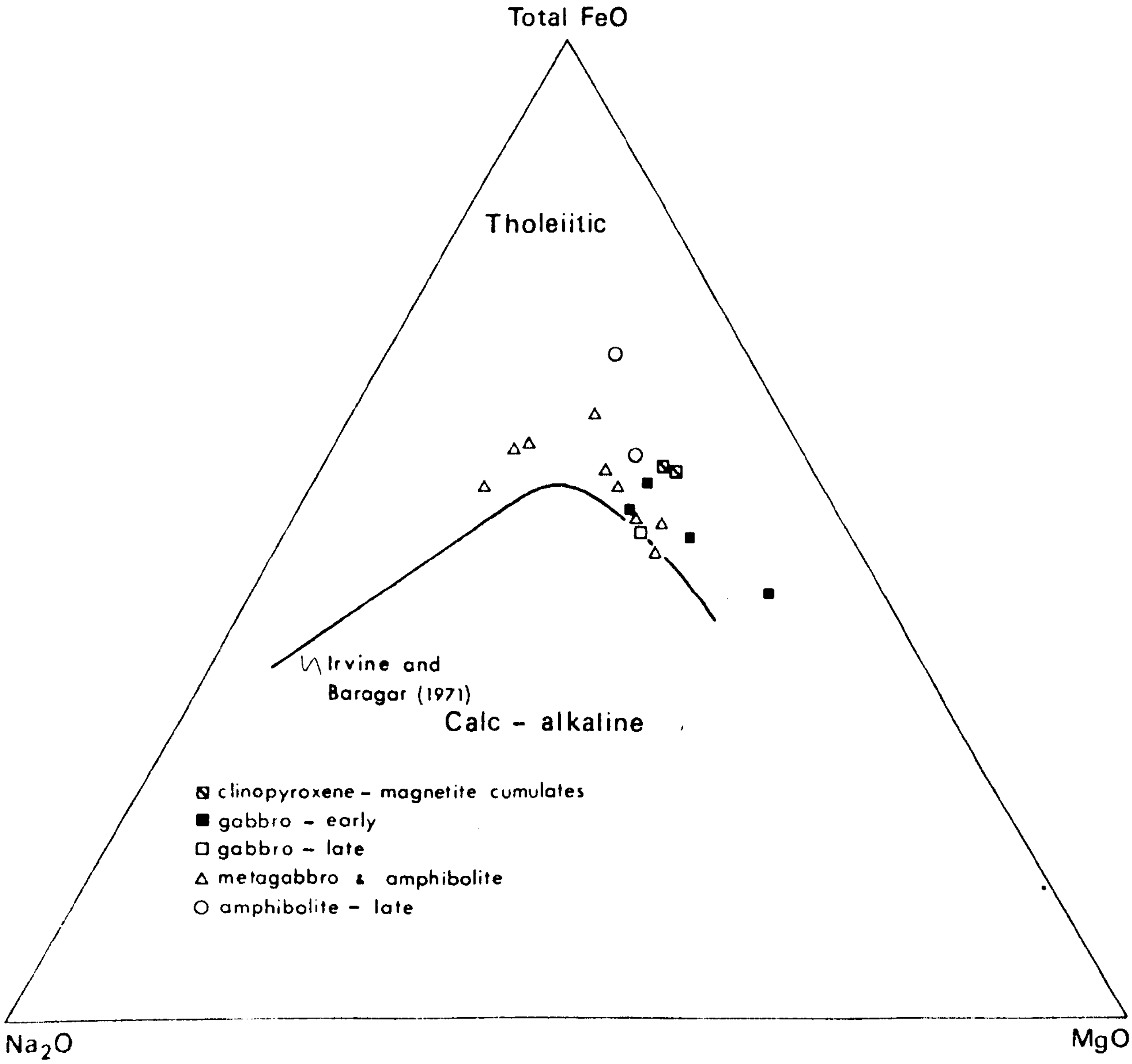
Problems with the use of major element data are twofold; firstly, there is a large overlap between the fields and secondly, many of the major elements are mobile, certainly during metamorphism and particularly under the influence of fluid activity. This casts some doubt on the use of many plots, especially those involving alkali elements.

The question of mobility of elements during metamorphism has been studied on gabbro-amphibolite pairs in South Norway by Elliott (1972) who indicated that metamorphism was not isochemical, with movement of K, Fe, Ca and probably Si, in the transition. Therefore, in the diagrams used, the essentially unaltered coronites were the only samples initially studied for trends. Garcia (1978) points out that the majority of analyses have been carried out on volcanic rocks and more work is necessary on plutonic suites. Since Petro et al (1979) have applied the chemical plots, used for volcanics by previous workers, to their plutonic rocks, the same techniques have been used in analysis of the Grøsli plutonics.

4.9 MAJOR ELEMENT ANALYSES OF THE GRØSLI 'VINOR' INTRUSIONS

Fig. 4.8 shows that the coronite analyses plot closer to the trend for 'extensional' rock suites than for 'compressional' suites, although there is some ambiguity at the higher CaO values where the two trends cross. An AFM plot (Fig. 4.9) shows that the coronites and their metamorphic equivalents fall within the tholeiitic field of Irvine and Baragar (1971). It should be noted, however, that this plot is incapable of distinguishing ocean floor tholeiites from island-arc tholeiites. There are insufficient data to give definite conclusions from the other diagrams used by Petro et al (1979) since extensive overlap between the trends requires a large number of points to define a line.

The relative positions of the metagabbros and amphibolites on the AFM diagram indicates that the metamorphism appears to have effected relatively little chemical mobility. This is not surprising on this diagram, since Elliott (1973) considered that total iron and magnesium remained static while potassium increased during amphibolitization. On the AFM diagram, such increases would lead to a slight movement towards the alkali corner of limited extent, since the absolute alkali percentage is low compared to that of the other two end members.



Total FeO
Tholeiitic
Irvine and
Baragar (1971)
Calc - alkaline
clinopyroxene - magnetite cumulates
gabbro - early
gabbro - late
metagabbro & amphibolite
amphibolite - late
Na_2O
K_2O
MgO

FIG. 4·9

Use of the discriminant function analysis diagrams of Pearce (1976), produced very wide scatters, with no patterns of distribution. The extent of scatter of these plots proves the limited applicability of such major element discriminations applied to rocks which have undergone metamorphism. The use of major elements in altered basic rocks is beset with problems created by the differential mobility of certain elements. At best, an extensional or compressional environment of formation may be suggested.

4.10 TRACE ELEMENT ANALYSES: RELEVANT PREVIOUS WORK

Trace elements have been widely used in determining basic rock protoliths. The so-called 'immobile elements' have been of special interest since they have been considered to retain their original concentrations during hydrothermal alteration and low grade metamorphism, e.g. Cann (1970), Pearce and Cann (1971, 1973), Bickle and Nisbet (1972), Wilkinson and Cann (1974). Floyd and Winchester (1978) used trace element data from spilites, Archaean greenstone suites and greenschist/amphibolite assemblages to define the origin of these rocks. The most commonly studied of these 'immobile elements' are Ti, Zr, Y, La, Ce, Nb, Ga and Sc.

In South Norway, Field and Elliott (1974) concluded that during the metamorphism of basic plutonic rocks to amphibolities, all the above elements, with the exception of Zr, remained static. Zr may have become slightly enriched within the amphibolites. However Pearce and Cann (1973) consider Zr to be essentially immobile during metamorphism.

4.11 TRACE ELEMENT ANALYSES OF THE GRØSLI 'VINOR' INTRUSIONS

Pearce and Cann (1973) used a plot of Ti versus Zr to discriminate between island-arc basalts (IAB), ocean floor basalts (OFB) and within-plate basalts (WPB). The plot utilised the fact that there is a general increase in Zr and Ti consecutively through the three groups above. Garcia (1978) confirmed this plot as one of the most useful for discrimination of island-arc and other magma types, but added that it was not always capable of discriminating between the two types when Ti contents were low. Fig. 4.10 shows that all of the Grøsli basic rocks plot well away from the IAB field and, whilst not actually lying completely within the MORB and WPB fields, certainly have Ti and Zr values consistent with ocean floor basalts.

Pearce and Cann (1973) also used a triangular plot of Zr/Ti/Y (Fig. 4.11) which also contained a field of overlap between volcanic arc lavas and ocean floor types. The very large scatter of the Grøsli rocks across all fields in this plot has been suggested (Pearce, pers. comm. 1980) to be due to fractionation and differentiation of the parent rocks (now coronites). This explains the rocks which plot towards the Ti apex, which are clinopyroxene-magnetite cumulates. The group of points which are encircled by dashes in Fig. 4.11 all plot within the WPB field and, with one exception, are coronites without cumulate textures (with olivine gabbro or gabbro mineralogies). These rocks are probably the parent rock type. The remaining and widest dispersed groups of rocks are all metagabbros or amphibolites. They could be explained as original derivatives of the parent magma along a fractionation trend. However, extreme caution must be exercised in reaching such a conclusion since the rocks have undergone hydration

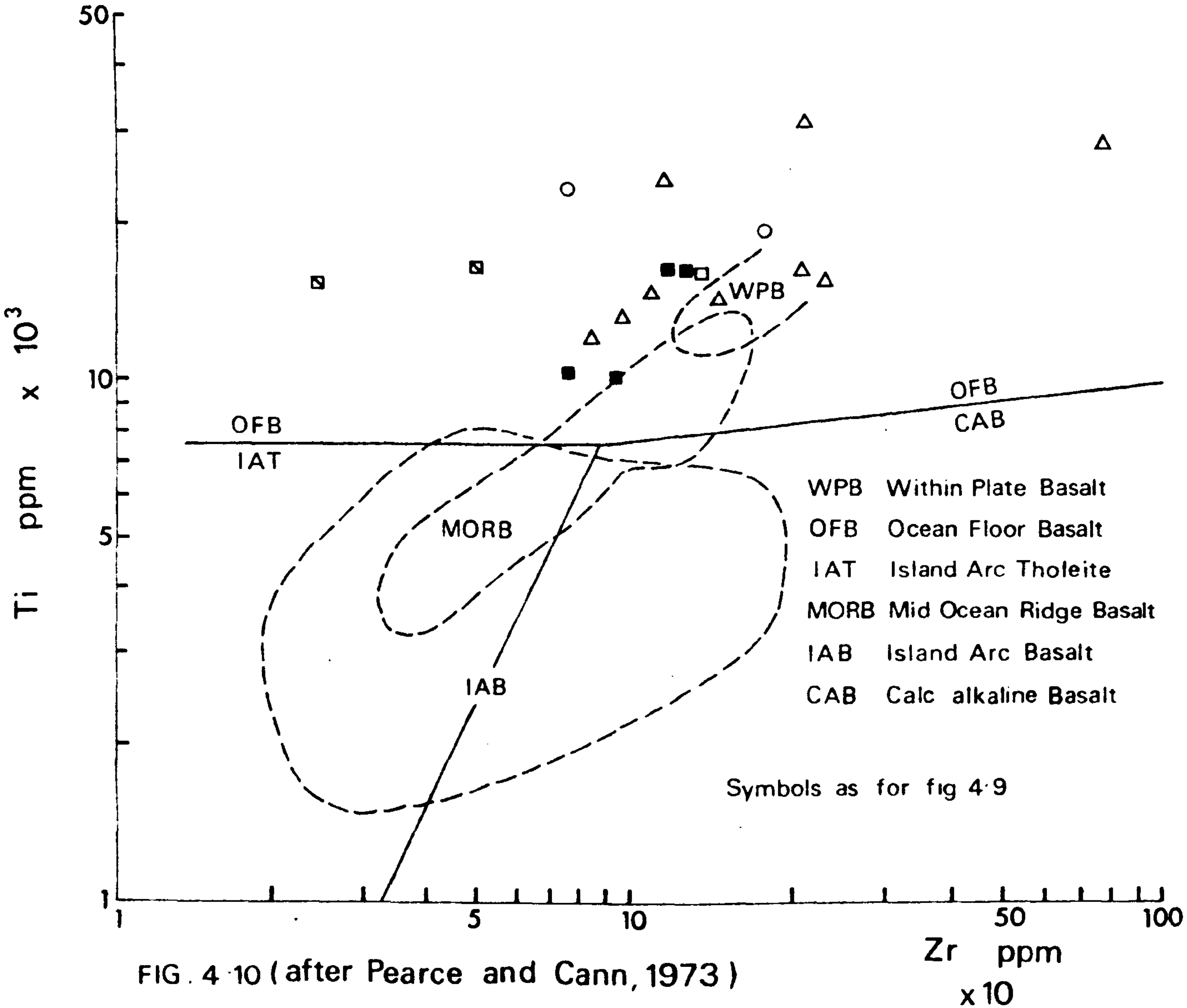


FIG. 4·10 (after Pearce and Cann, 1973)

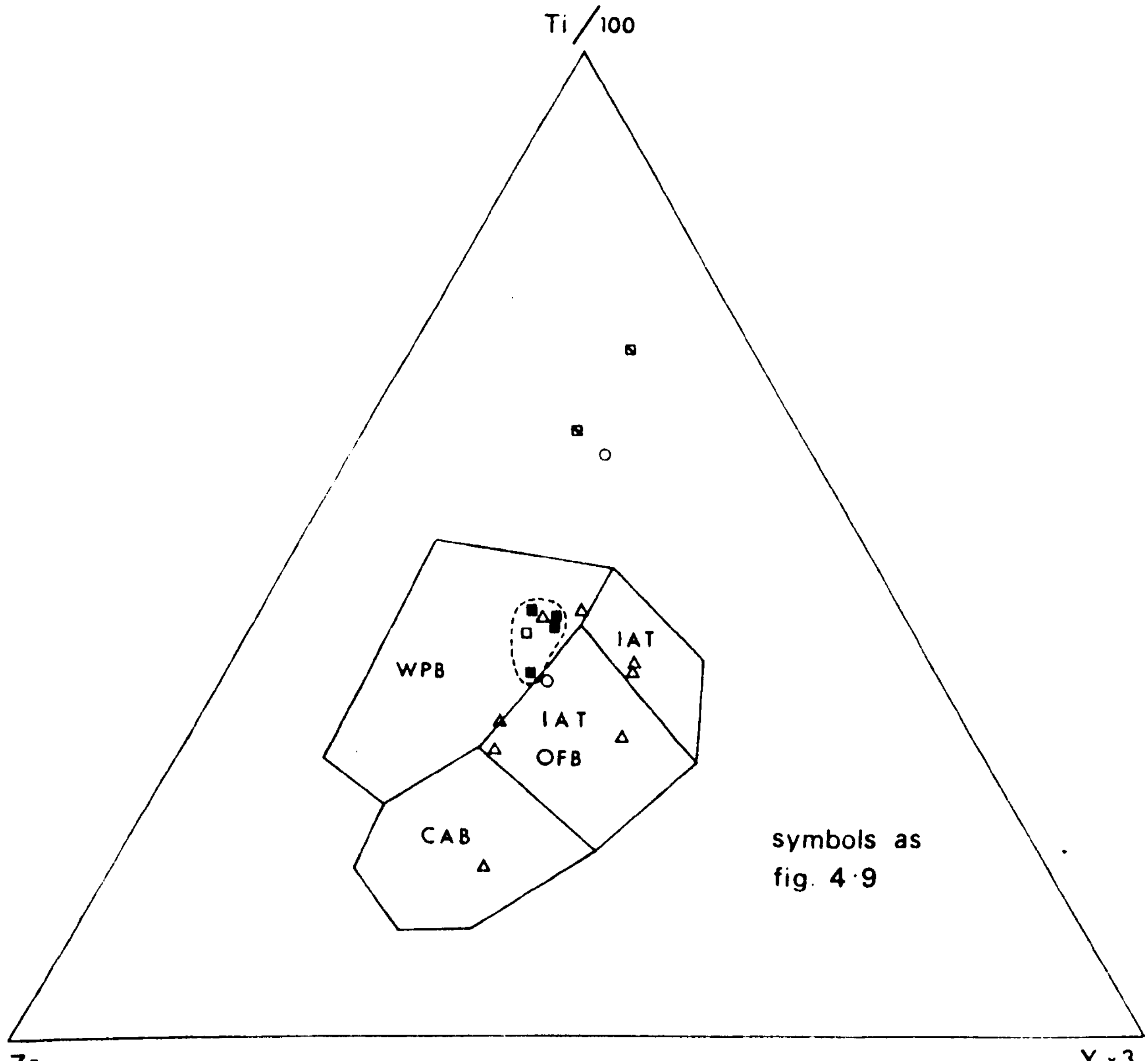


FIG. 4·11 (after Pearce and Cann, 1973)

during metamorphism (in some cases extensive) and it is not clear from this diagram whether magmatic or metamorphic fractionation is responsible for the trend.

Pearce and Gale (1977) used a plot of Zr/Y versus Ti/Y to discriminate between WPB and plate margin basalts (PMB). This diagram makes use of the enrichment of Zr and Ti but not Y, characteristic of within-plate basalts. The plot effectively removes the problem of overlap between WPB and mid-ocean ridge basalt (MORB) in the Ti versus Zr diagram (Fig. 4.10). The same authors also used, in combination with the above plot, a discriminant analysis based on the fact that for any Ti value, the Cr value of island-arc basalts (IAB) will be lower than for ocean floor basalts (OFB and MORB). These two plots thus split the three fields of IAB (calc-alkaline and tholeiitic), OFB (MORB) and WPB without any overlap.

A Zr/Y versus Ti/Y diagram for the Grøsli rocks (Fig. 4.12) shows a spread of points across the PMB/WPB boundary. The coronites plot mainly within the WPB field, but very close to the boundary, while the cumulate differentiates plot well within the WPB field by virtue of their high Ti content. The metagabbros and amphibolites plot largely within the PMB field; and if these rocks are then plotted on a Cr versus Ti diagram (Fig. 4.13) it can be seen that they fall well within the MORB field. On the basis of these two diagrams it seems therefore that the rocks may represent either: (i) within-plate basalts (WPB) which have had their chemistry somewhat altered during metamorphism to resemble MORB types, or (ii) two magma types, or (iii) one transitional WPB/MORB magma type. Fractionation processes might be expected to have enhanced any original differences by virtue

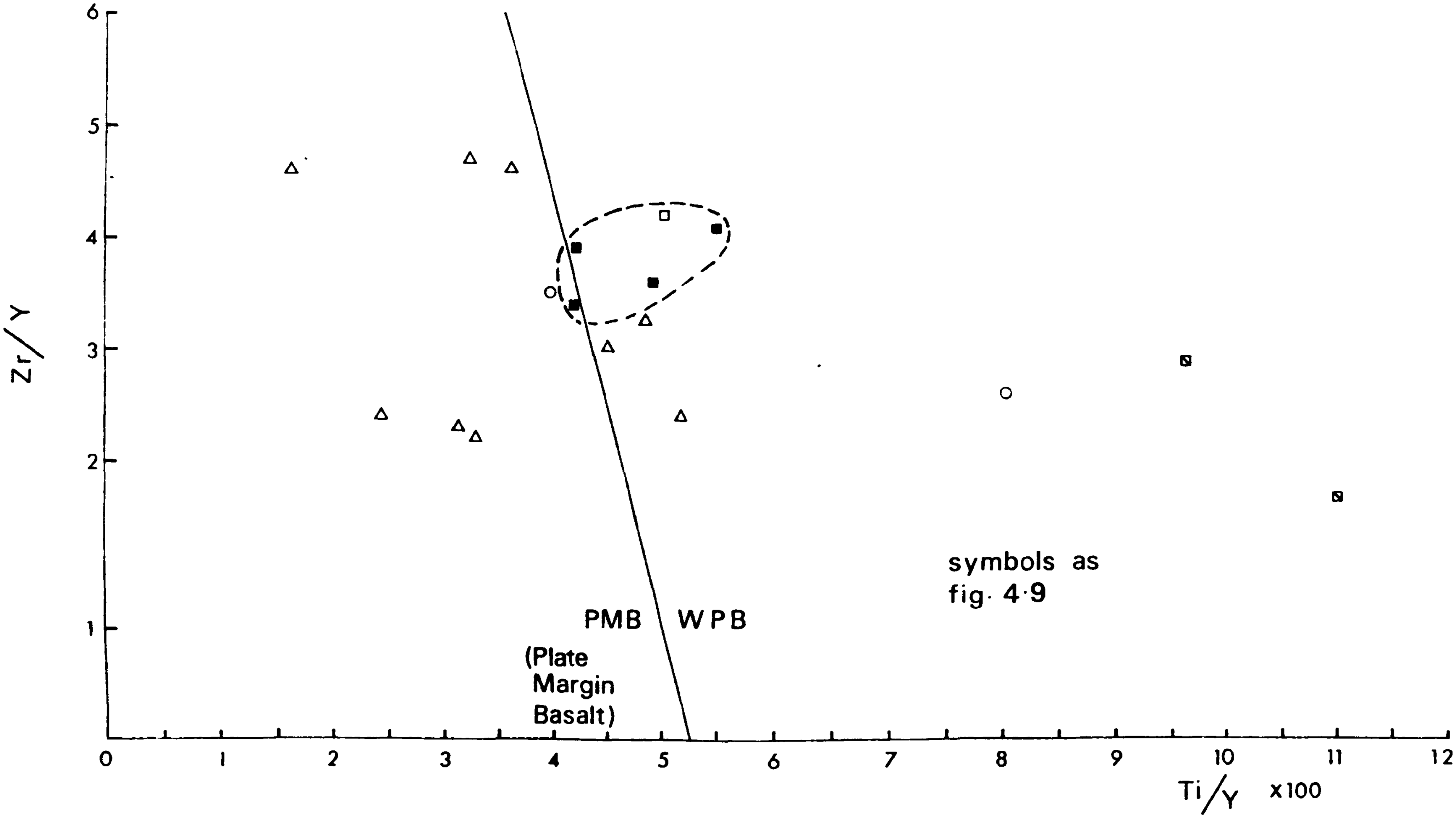


FIG. 4·12 (after Pearce and Gale, 1977)

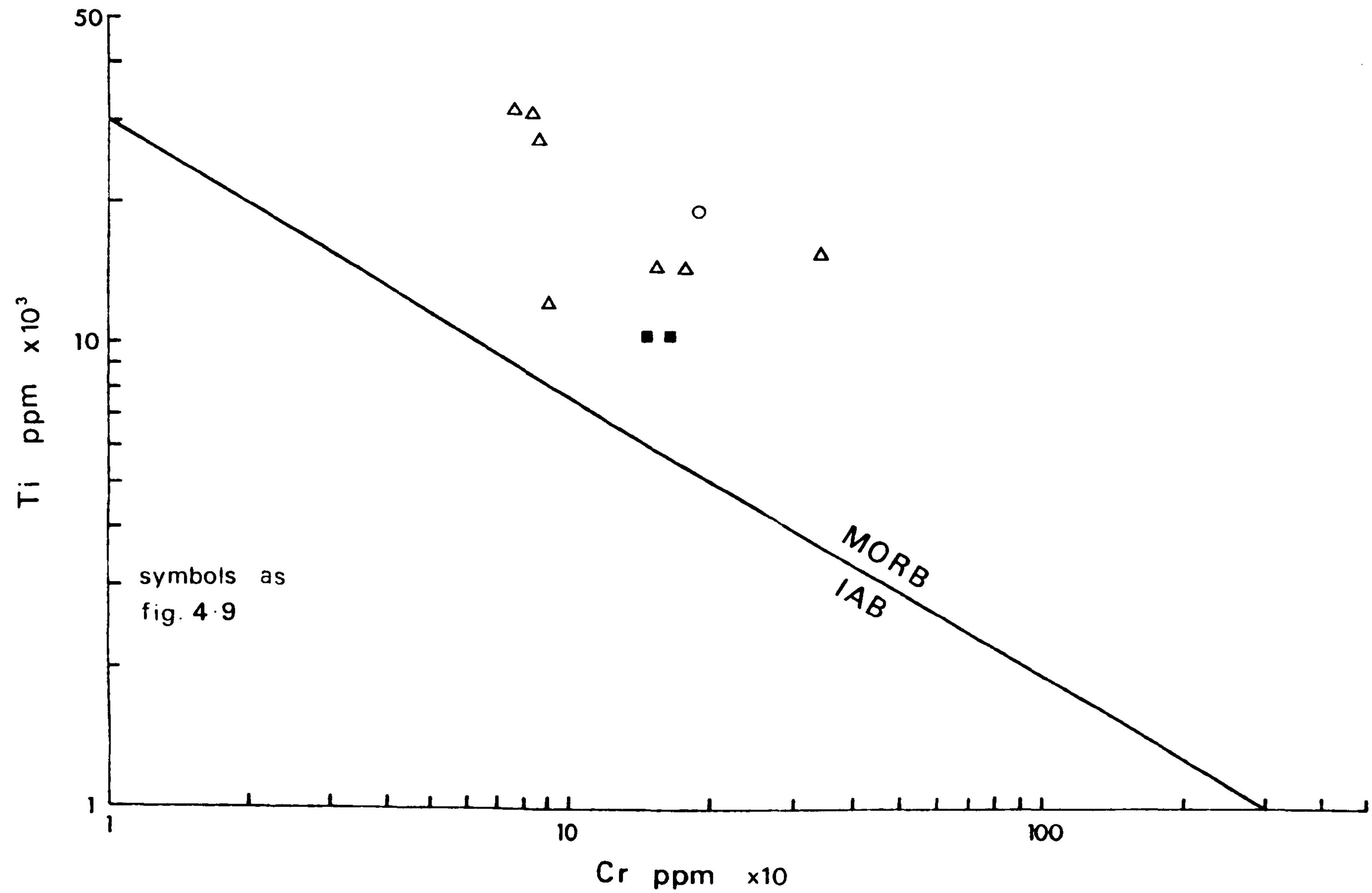


FIG.4·13 (after Pearce, 1975)

of concentrating the incompatible elements.

Pearce (in press) has used a number of discriminant plots to determine environments. These plots included Zr/Y versus Zr, Cr versus Y and Cr versus Ce/Sr. The first of these plots (Fig. 4.14) produces a trend very similar to Fig. 4.11. The majority of points plot in the MORB field with some in the WPB field. A diagram of Cr versus Y (Fig. 4.15) produces effectively the same result. A Cr versus Ce/Sr diagram has not been used in the Grøsli study since there is evidence (Field and Elliott, 1974) that Sr was fairly mobile in the Bamble Series rocks of South Norway and this may also have been the case in the Kongsberg Series.

Several common factors emerge from the discriminant plots:

1. There is a group of analyses (mainly of coronites; ringed in Figs. 4.11, 4.12, 4.14, 4.15) which are always close to the WPB-MORB boundary and often traverse it.
2. A group of magnetite-clinopyroxene cumulates plot in areas of higher Ti and somewhat lower Zr values. Their cumulate nature causes them to plot in various fields ranging from IAB through OFB to WPB depending on the diagram used.
3. A group of metagabbros and amphibolites plot in both MORB and WPB fields, but with a preponderance in the former.

The variations described above between the coronites and the more metamorphosed rocks can be ascribed to one or more possible causes:

a) Igneous fractionation.

b) Metamorphism.

c) Two distinct magma types.

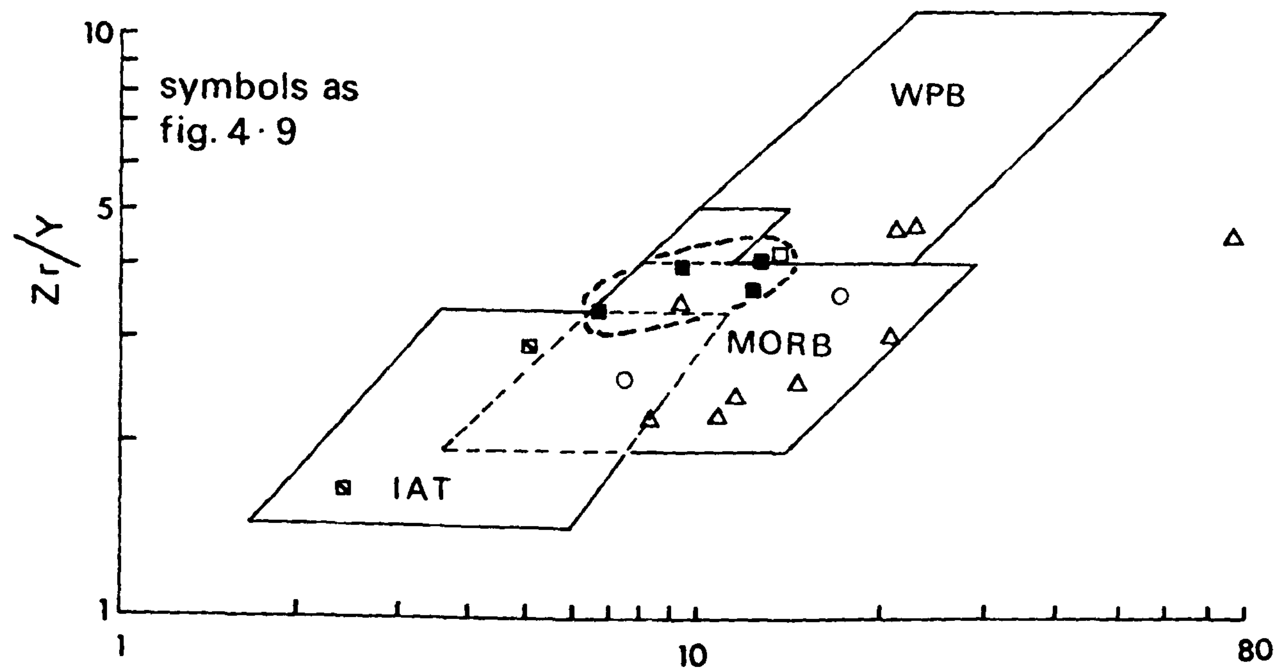


FIG. 4·14 (after Pearce, in press)

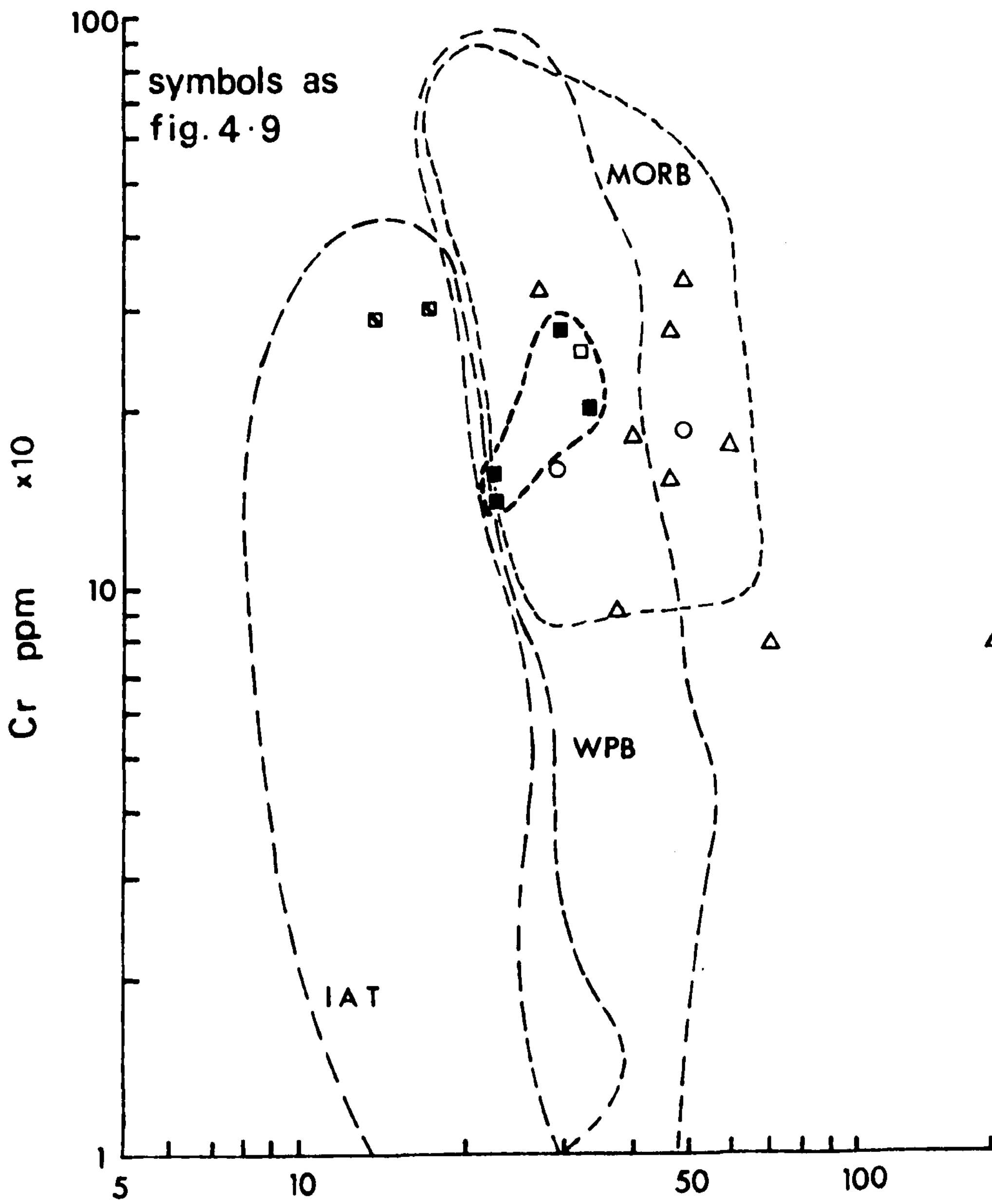


FIG. 4·15 (after Pearce, in press)

d) One magma of transitional WPB/MORB character.

The field relationships argue against the metagabbros and amphibolites being genetically distinct from the coronites (i.e. formed from a separate magma). While extreme igneous fractionation could cause the differences in the location of the plotted points, it is not a likely event in the Grøsli area since all the rocks are of basaltic composition. Pearce (in press) notes that even acid end members of basaltic fractionation trends have tended to remain in their parental fields. However, the differentiated cumulates of the Grøsli area do frequently plot in separate fields. Thus, the combined effect of metamorphism and fractionation may well have enhanced any differences, but these two factors alone are considered unlikely to have caused the entire variability, particularly since these elements are usually considered immobile.

Pearce and Gale (1977) recognized a transitional rock type showing both OFB and WPB characteristics. They cited three Norwegian examples of basaltic hosts associated with ore deposits [Joma (Grong) and Røros, both of Caledonian age and Bidjovagge (Finnmark), of Precambrian age]. They considered that these rocks were formed at a continental margin in a small ocean of Red Sea type. Pearce (in press) has stated that those transitional magmas are characteristic of diffuse spreading centres (such as occur today in Iceland, Afar and the Azores) rather than at normal ocean ridge locations.

The major part of the Grøsli 'Vinor' body is formed from a single intrusive phase and its trace element characteristics can be explained largely in terms of it being of a transitional nature of both MORB and WPB types. If analogies are feasible between Precambrian environments and Mesozoic-Cainozoic plate-tectonic regimes, then the Grøsli 'Vinor' would appear to be the deep level equivalent of lavas erupted at a divergent plate margin. The rocks were intruded into continental crust (the supracrustals) in a similar environment to Red Sea and Afar-type spreading and thus represent the commencement of rifting (i.e. an incipient constructive plate margin). The partial WPB characteristics of the early magma would thus be explained in terms of this being a juvenile rather than a mature rift. That this rifting was aborted (possibly in a similar way to the East African Rift system) is shown by the fact that the Sveconorwegian orogeny occurred relatively soon after the 'Vinor' intrusions.

CHAPTER FIVE

THE ORE DEPOSIT AT GRØSLI AND ITS METAMORPHISM

5.1 INTRODUCTION

At Grøsli, both the massive ore and the disseminations consist of pyrrhotite, pyrite, sphalerite and chalcopyrite. Any of the four phases may be dominant although sphalerite and chalcopyrite are the least abundant and may be entirely absent in some specimens. Galena is present in accessory amounts. Oxides (magnetite, ilmenite and rarer rutile) are present within the sulphide assemblage where the ore is enclosed in basic rock. The ores within the gneisses are oxide free (except for supergene alteration). Within the basic rocks, the oxides may form up to 20% of the ore assemblage but usually comprise less than 5%. Ilvaite also occurs as very rare grains, again only within ore enclosed by basic rock. Gangue minerals include sericite, chlorite, actinolite, quartz and calcite.

The ore bodies at Grøsli underwent high grade regional metamorphism producing recrystallization and annealed textures whilst non-synchronous directed stress resulted in deformation. The ore deposit can be split into two groups: ore with complex textural features (at the mine and in a northerly extension) and ore with less complex features (to the south).

5.2 THE MINE ORE BODY

Several fabrics have developed within what is the northernmost part of the ore deposit. The differing textures are a result of the mineralogy of the assemblages since the difference in the competence of the minerals has caused them to react differently under the same metamorphic conditions. The fabric now visible is due largely to a late brittle stress which produced a gneissic fabric in the ore. The other dominant feature is the presence of large, early pyrite porphyroblasts which (where present) have prevented the development of the gneissic texture.

5.2.1 Mineralogy

Very large pyrite porphyroblasts occur in a matrix of pyrrhotite with minor sphalerite and chalcopyrite. Perfect pyrite cubes have edges up to 15 cm long and reflect a static growth period. They contain inclusions of all the other sulphides and are often broken, the cracks usually being infilled by chalcopyrite or silicate gangue or an intergrowth of the two (Fig. 5.1).

Pyrrhotite exhibits only xenoblastic habits and is commonly intergrown with sphalerite and chalcopyrite. Pyrrhotite crystals show both early pre-tectonic thermal recrystallizations and later deformational overprints. Very pronounced glide twinning, kink banding and undulose extinction are common (Figs. 5.2, 5.3). Annealed features include equidimensional mosaics of crystals often showing 120^{o} triple junctions (Fig. 5.4). Sub-grain development (Fig. 5.5) is also common. Crystals also occur exhibiting no strain.

Sphalerite and chalcopyrite are usually both minor phases. The sphalerite crystallized at the same time as the pyrrhotite although it behaved somewhat differently under stress. Almost all sphalerite crystals contain extensive exsolution blebs of chalcopyrite which occur in strings

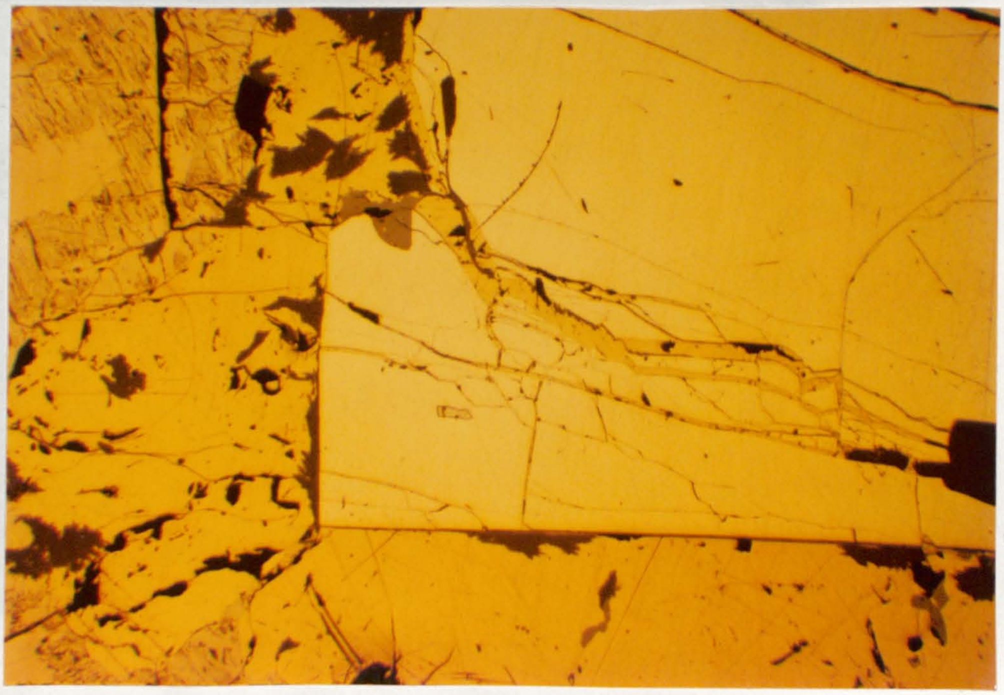


FIG. 5.1. Chalcopyrite and gangue infilling brittle cracks in pyrite. Plane polarized light (PPL). Oil emersion. X64.

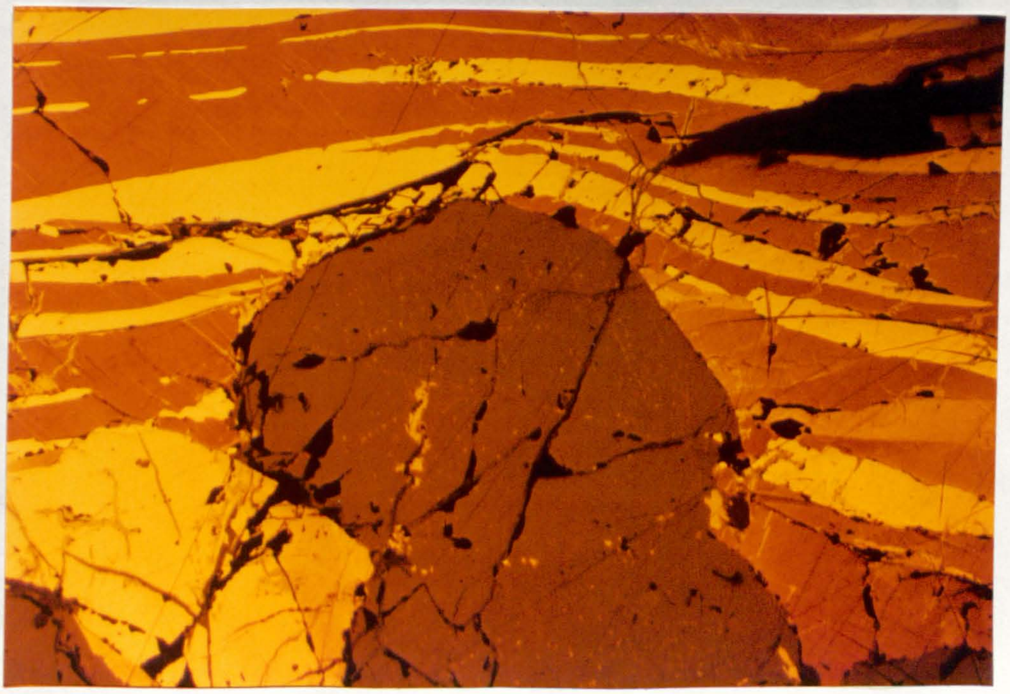


FIG. 5.2. Glide twins in pyrrhotite flexed around sphalerite. Crossed polarised light (XPL). Air. X25.

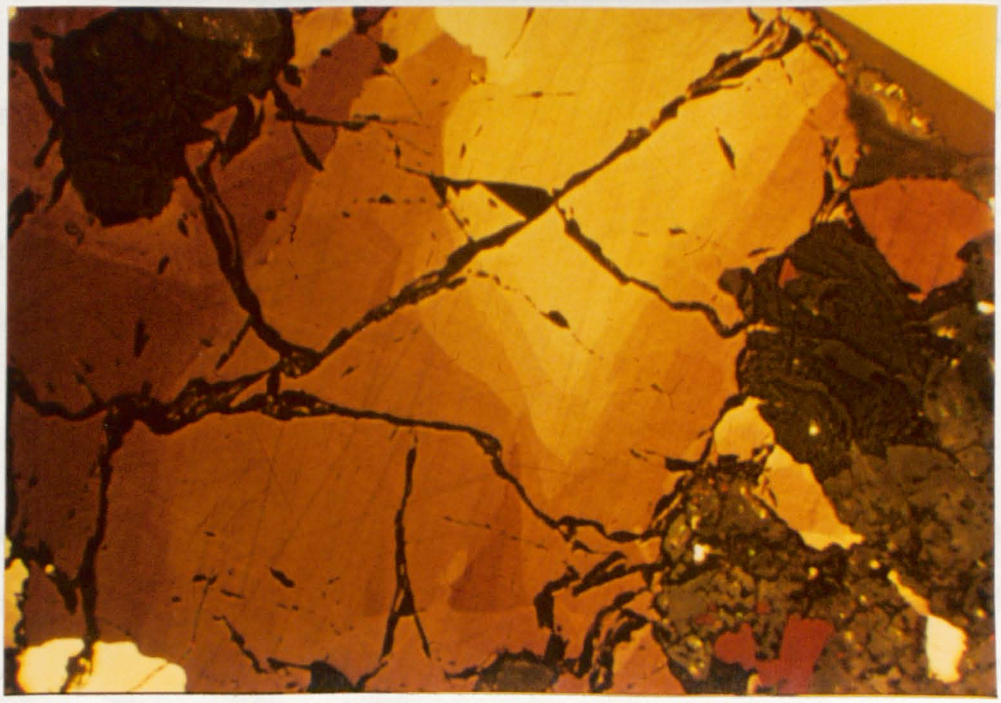


FIG. 5.3. Strain extinction in pyrrhotite grains. XPL. Air. X10.

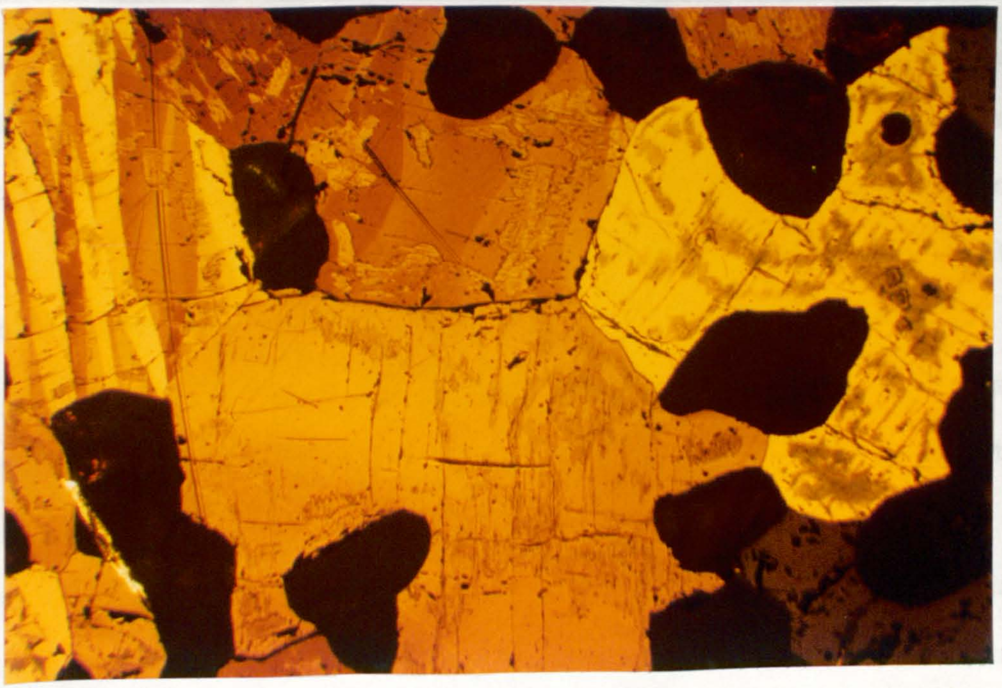


FIG. 5.4. Equilibrium triple junctions in pyrrhotite crystals. XPL. Air. X10.

along the crystallographic directions. Chalcopyrite often exhibits intrusive and replacive textures in the other sulphides and is often associated with silicate gangue. The fact that chalcopyrite infills even the brittle cracks in pyrite indicates that it was still mobile late in the history of the ore body.

5.2.2 Fabric

Where little or no pyrite is present, a strong, coarse gneissic texture has developed. Compositional banding (possibly primary) is fairly common on a microscopic scale and zones of pyrrhotite-sphalerite can be distinguished from zones of dominantly pyrrhotite composition (Fig. 5.6). The latter bands exhibit a range of textures. Pyrrhotite glide twins are highly distorted and in places shear has continued and produced fracture of the crystal lattice. Strong undulose extinction is also common. However, textures of annealing have been overprinted on this highly sheared fabric (Fig. 5.7). Since the stress effects are still visible, it is clear that the recrystallization was incomplete. Glide twins in pyrrhotites outside the pyrrhotite bands become polygonized at their boundaries as they approach these recrystallized bands (Fig. 5.8). These bands are the zones along which the most extensive movement and deformation has occurred. Stanton (1972) states that "under a given set of conditions, that portion of a grain or of a single phase aggregate that has been deformed most severely will crystallize most rapidly". Extensive sub-grain development also occurs within these bands and Stanton again considers that this type of texture could be due to a partial annealing. These bands also contain highly xenoblastic crystals with ragged margins and strong undulose extinction. These

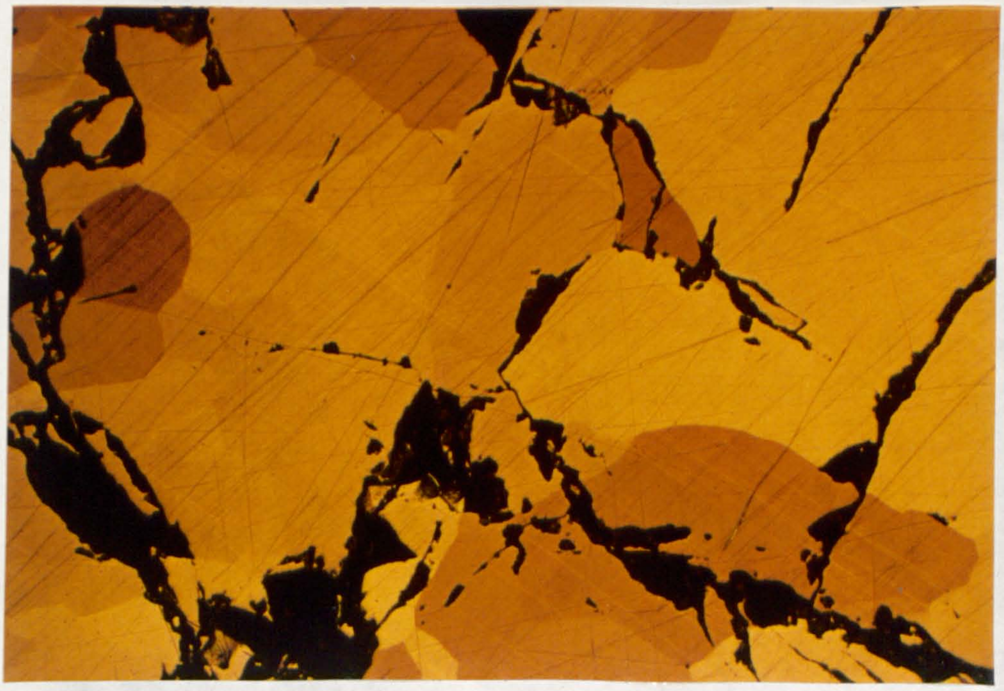


FIG. 5.5. Sub-grain development in pyrrhotite. XPL. Oil. X64.

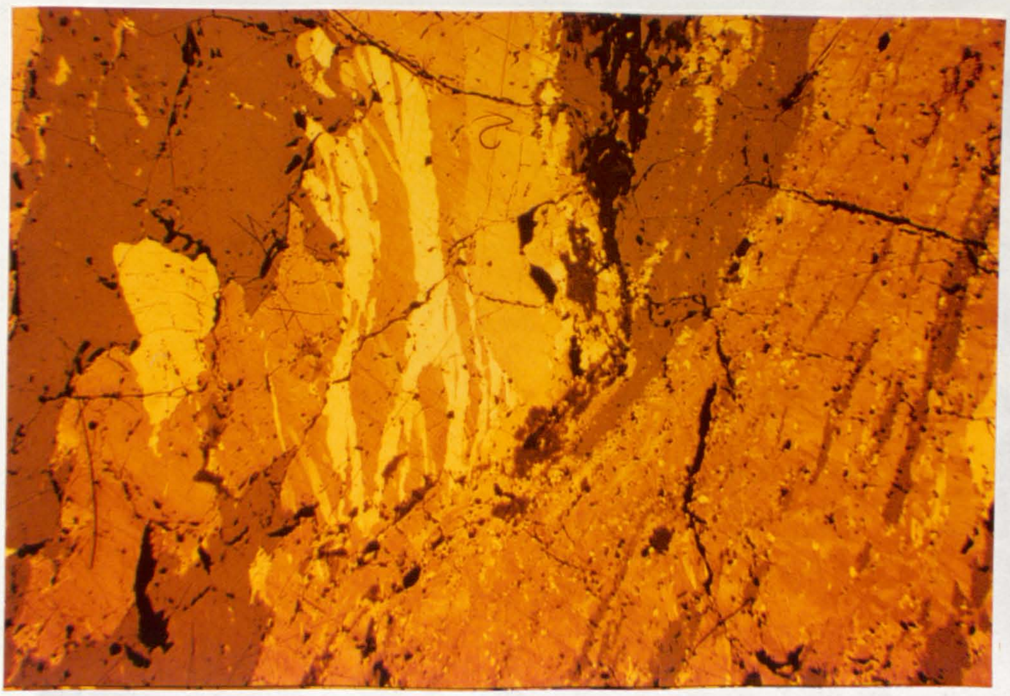


FIG. 5.6. Junction of highly sheared, pyrrhotite dominated zone (to the right of the photo) and less sheared pyrrhotite-sphalerite zone (to the left). XPL. Air. X10.

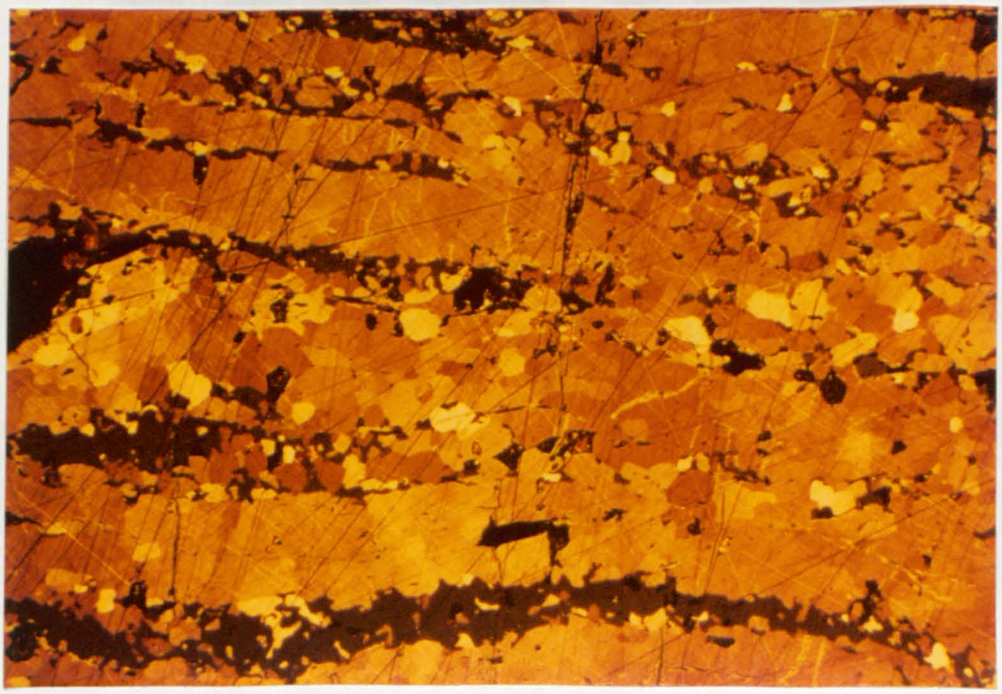


FIG. 5.7. Highly sheared pyrrhotite zone exhibiting post-tectonic re-crystallization to a mosaic of unstrained crystals. XPL. Air. X64.

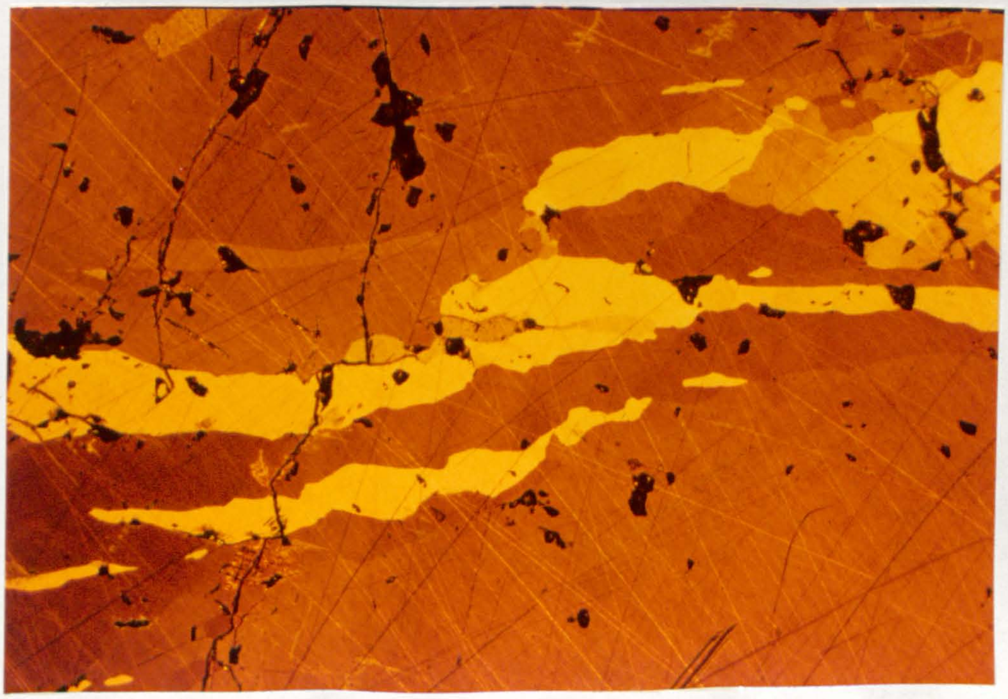


FIG. 5.8. Polygonization (recrystallization) of deformation glide twins in pyrrhotite. XPL. Oil. X160.

crystals are larger than the recrystallized matrix and are often rounded or exhibit augen shapes. They are pyrrhotites which have not recrystallized and represent relics of the fabric produced by the deformation (Fig. 5.9). Further evidence of the high degree of shear within the pyrrhotite bands is provided by the occasional sphalerites present. They have become very elongated and drawn out in flasers parallel to the shear direction.

Within the pyrrhotite-sphalerite bands, the two minerals are present in equal proportions. The pyrrhotite crystals are much larger than in the pyrrhotite rich bands. Several types of grain junction can be identified. Rare, relict triple junctions occur (Fig. 5.4) but more usually junctions are in high energy states: kinked, cuspate or ragged and broken junctions are most common. An earlier equilibrium mosaic of crystals has been overprinted by the gneissic texture during deformation. The gneissosity is not so well developed in these bands and the pyrrhotites have maintained their shape, showing only glide twinning and kink banding without rupture. Clark and Kelly (1973) state that at 250°C pyrrhotite loses its strength, becomes ductile and twin gliding commences. The sphalerites present have thus acted as resistant masses against which the more ductile pyrrhotite has been deformed (Fig. 5.10). The presence of the sphalerite has prevented the excessive shearing of the pyrrhotite component. Stress cannot have operated over a long period of time since the pyrrhotites have not flowed by creep around the sphalerites. It is suggested that the pyrrhotites were squeezed against the sphalerites rapidly rather than slowly. Two twin directions occur within the pyrrhotites in this band type. Clark and Kelly state that these develop with one set at a high

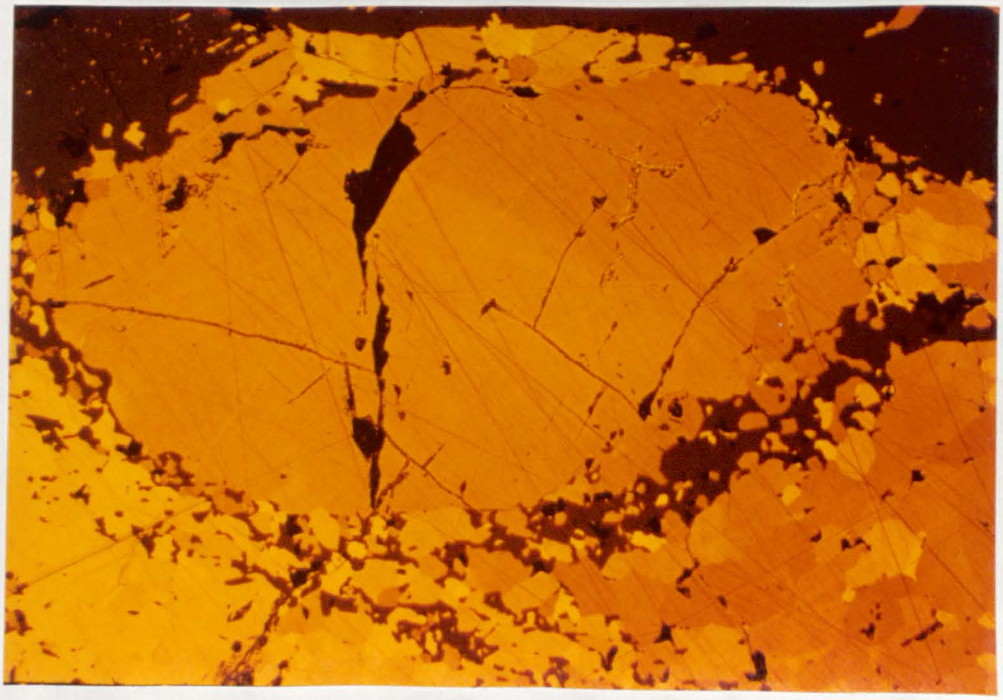


FIG. 5.9. Relict, syn-tectonic augen of pyrrhotite within a matrix of post-tectonic recrystallized pyrrhotite. XPL. Oil. X64.

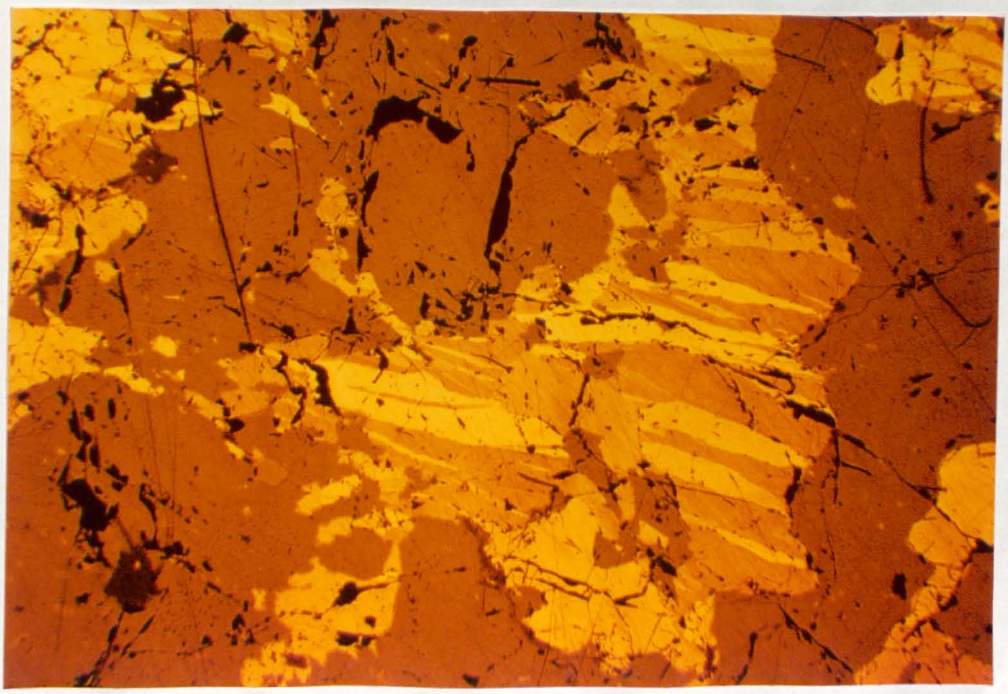


FIG. 5.10. Pyrrhotite-sphalerite zone exhibiting strong glide twinning in the pyrrhotites. XPL. Air. X10.

angle to the maximum compression direction and one set at a low angle.

No later features of annealing occur within these latter bands, suggesting that the recrystallization of the pyrrhotite bands is due to either frictionally generated heat or due to the high degree of internal energy possessed by the highly deformed crystals. Where pyrite is dominant, the gneissic textures are even less well developed and, in the zones containing the very large porphyroblasts, there is no foliation. Early triple junctions are very well developed between crystals within the vicinity of the pyrites. Strong glide twinning occurs in the pyrrhotites where they have been deformed against the pyrites which have acted as resistant blocks. The same stress has caused the cracking of the pyrites. Pyrrhotites surrounded by pyrite cubes lack stress features due to the protection afforded them by the competent pyrites (Fig. 5.11). The pyrites have grown to their large size because of favourable conditions. Firstly, the matrix in which they occur is dominantly pyrrhotite which itself has a low crystalloblastic energy. Secondly, the combination of the heat from the metamorphism and the huge gabbro body provided high temperatures with a slow cooling rate to allow the pyrite time to grow to substantial sizes.

Ores from borehole 4 also exhibit the same fabrics found in the mine. Some of the pyrites present are due to supergene 'birds-eye' alteration (Fig. 5.12) but this is not a common features. Other ('dust-bin') pyrites contain much sphalerite especially towards their margins. The original pyrite may have contained many impurities (suggesting a non-magmatic origin) which on metamorphism were mobilized and migrated to the crystal edges where they coalesced into larger grains.

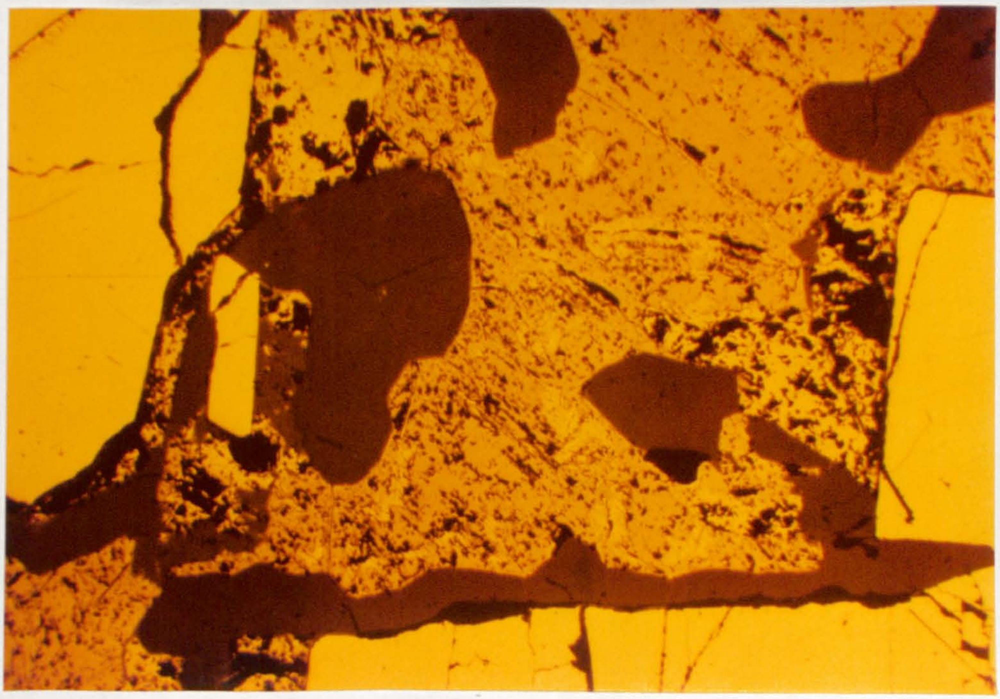


FIG. 5.11. Unstrained pyrrhotite cyrstals protected by pyrite cubes.

XPL. Air. X10.

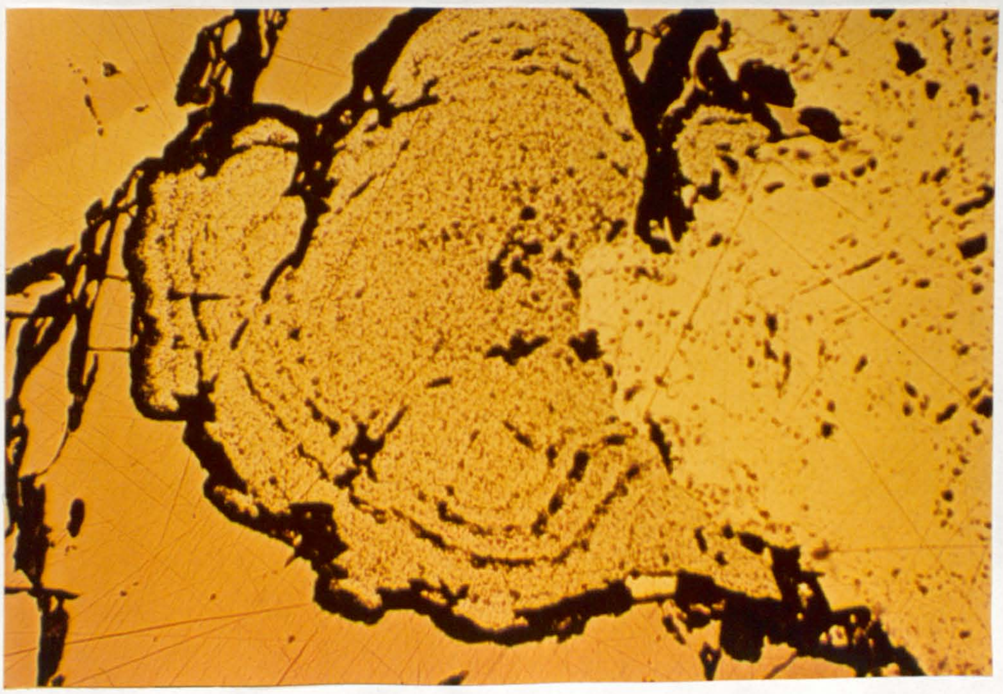


FIG. 5.12. Supergene 'birds-eye' alteration of pyrrhotite to pyrite.

PPL. Oil. X64.

5.2.3 Deformation and recrystallization history

The size of the pyrite porphyroblasts and their well developed shapes together with the relict polygonal mosaic of large crystals, with good equilibrium triple points, suggests an early phase of high temperature recrystallization of the ore body. Over this fabric a deformational episode has produced a gneissic texture. Glide twins and undulose extinction (sometimes superimposed over crystals still exhibiting relict triple point junctions) have developed together with ragged and irregular disequilibrium crystal boundaries. Cracking of the pyrite cubes is also a feature of this phase. Deformation was low grade with little or no annealing later, except in the most highly deformed pyrrhotite-rich bands of the gneissic ore. Finally, chalcopyrite flowed into cracks produced in the pyrite. Within the mine area a late brittle shear plane dips to the east, but is not found to the south. The stresses which caused this shear plane were also responsible for the deformational phase which produced the gneissic texture of the ore and the brittle fracture of the pyrite.

5.3 THE SUB-SURFACE ORE BODIES (SOUTH OF THE MINE)

Sulphides occur in both gneisses and basic rocks, in some cases within distinct veins. However, in most cases within the basic rocks, the ore has a late-stage interstitial appearance and is usually totally xenoblastic where associated with silicates (Fig. 5.13). Where the ore is massive, it shows much less severe deformation features than the mine ore. The sulphides occur in interstitial relationships with unaltered coronite in borehole 10. In borehole 11, sulphides and micas crystallized

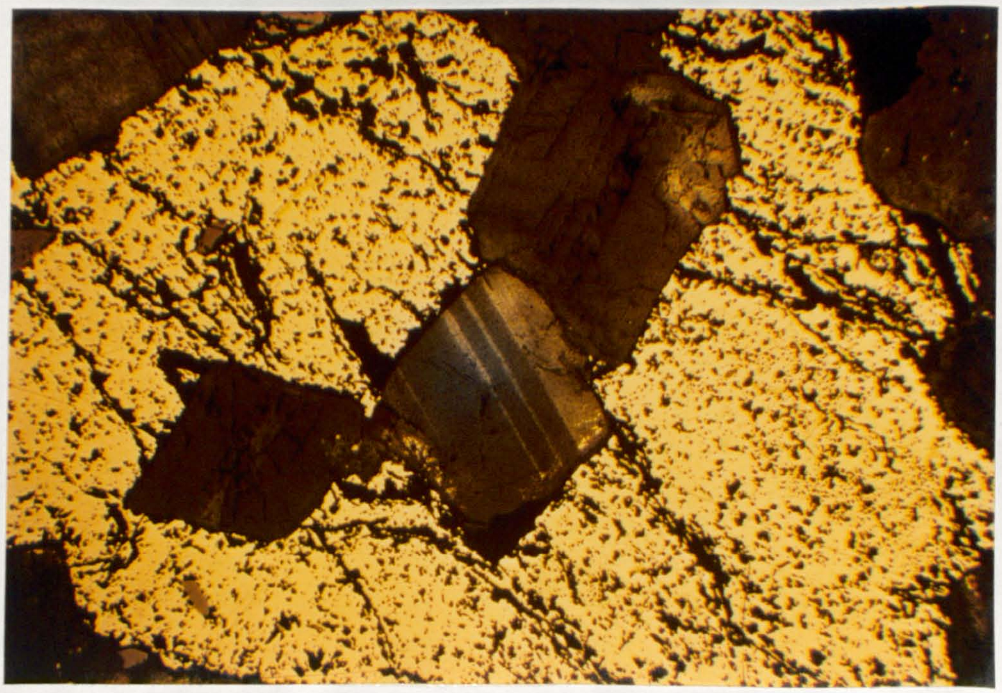


FIG. 5.13. Interstitial pyrrhotite surrounding plagioclase laths in coronite. XPL. Air. X10.

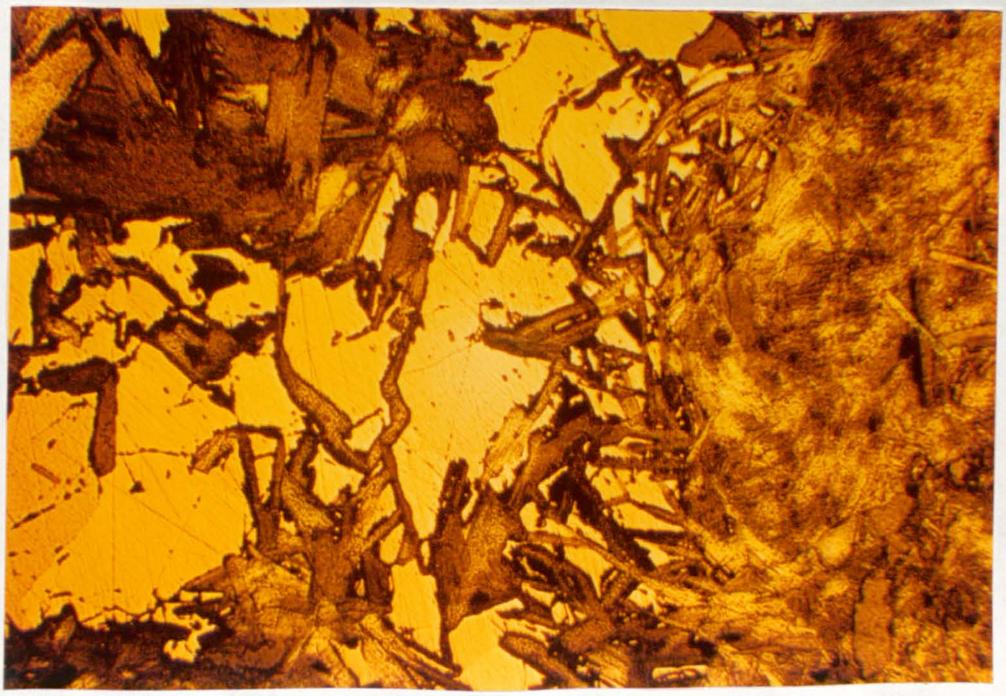


FIG. 5.14. Synchronously crystallized pyrrhotite and micas in hydrothermally altered gabbro. XPL. Air. X10.

synchronously to form a total replacement and hydration of the original basic rock (Fig. 5.14). To the north of borehole 10, the massive developments are located within gneisses.

Crystal size in these massive ores is generally greater than in the mine ore, especially where they are enclosed by basic rock. Pyrite porphyroblasts are developed throughout the Grøsli deposits, but never reach the sizes of those present in the mine. Ores from borehole 11 have no deformed crystals but those from 7 and 10 show some strain with undulose extinction and rare twinning (although effects are much weaker than in the mine samples). Deformation features become more common and more strongly developed towards the margins of the ore bodies. Silicate and oxide inclusions within the sulphides show partial sulphidization (Fig. 5.15). Ductile sulphides show disequilibrium boundaries with strongly serrated junctions resembling corrosive replacement features. Most silicate and oxide crystals show early stages of replacement with zones of sulphide within them. Chalcopyrite generally consolidated later than the other sulphides and shows strong replacive characteristics against all other minerals.

The sulphides thus appear to have reacted with the silicates and replaced them. Pyrite and pyrrhotite occupy large areas where they have totally replaced silicates. In other places the sulphides (especially pyrite) poikiloblastically enclose oxides and silicates. The oxides are commonly rounded in shape suggesting they have been caught-up, rolled and enclosed within the sulphides (Fig. 5.16). Ilmenite and magnetite both commonly occur either as individual grains or as a patchwork intergrowth of the two. Ilmenite exsolution from magnetite and rutile exsolution from both ilmenite and magnetite are also present. Massive

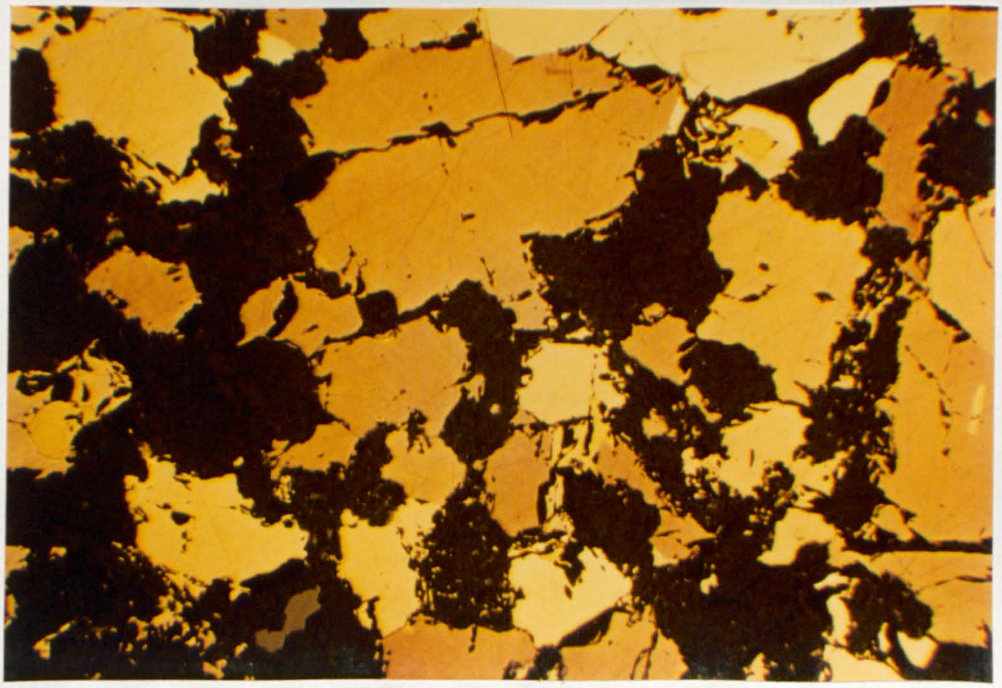


FIG. 5.15. Sulphidization of silicate host rock (gabbro). PPL. Air. X10.

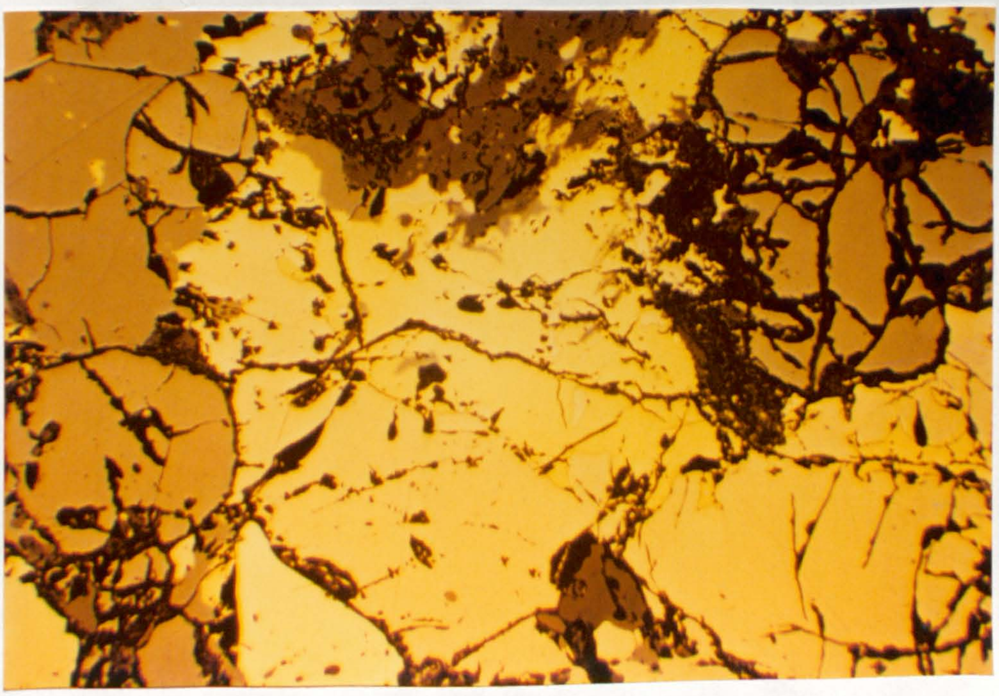


FIG. 5.16. Rounded magnetite xenoblasts in sulphide matrix. PPL. Air. X10.

sulphides within the gneisses generally show more intense deformation features than those within basic rocks.

5.3.1 Deformation and recrystallization history

It is suggested that the ore within the coronites and metagabbros represents material that was caught-up and melted during the intrusion of (anhydrous) olivine gabbro. Pyrrhotite (M.P. 1190^{o}C) is common within these rocks. Whilst the gabbro may have reached this temperature when totally molten, the early crystals (olivine, magnetite and calcic feldspars) appear to have already formed at the time the ore was incorporated. Therefore, it is possible that the pyrrhotite formed as a result of the incongruent melting of pyrite which occurs at 742^{o}C. Recrystallization of the sulphides resulted in the development of interstitial textures. Some of the ore present in borehole 11 is associated with hydrothermal alteration products of the original silicates and has crystallized synchronously with the hydrous minerals (mainly chlorite but also sericite-muscovite and some calcite). The re-mobilization of the ore into the rock in this case is obviously associated with localised hydrothermal activity and cannot be traced to any other borehole.

A suggested mode of formation of the hydrothermal ore is as follows. The ore may have been re-mobilized by fluids during greenschist facies metamorphism, after the intrusion of the gabbros, and after the epidote-amphibolite overprint, since ore occurs surrounding minerals formed during this stage in 'low grade' amphibolites. Where sulphides occur across the gabbro-gneiss junction (e.g. borehole 11), the boundary can be determined by the occurrence in the sulphide assemblage of oxides which can only have come from the basic rock. In borehole 11, as in many of the other boreholes at Grøsli, there is a clear but gradational succession towards the gneiss margin, reflecting progressive hydration from coronite through metagabbro to amphibolite. The hydrothermal ore deposits are present at the periphery of the basic intrusions located in rocks which have undergone extensive hydration. The sulphide-mica assemblages often exhibit relict sub-ophitic textures indicating the original igneous parent rock. These factors therefore suggest a post-intrusion re-mobilization and relocation under hydrous conditions of a minor part of the ore deposit.

5.4 AGE CORRELATIONS BETWEEN ORE AND HOST ROCK METAMORPHISM

The presence of sphalerite and lack of nickel bearing sulphides in the ore assemblage indicates that the ore was not of magmatic origin and may have been part of the supracrustal sequence. This sequence underwent amphibolite facies metamorphism in Svecofennian times. The ore would thus have a thoroughly recrystallized fabric prior to the development of the textural features described above. The Sveconorwegian orogeny reached mid-amphibolite facies conditions. This, and the intrusion of the Vinor basic bodies at the start of the metamorphism, led to a total recrystallization of the ore body. If oriented fabrics were developed by the stresses of the gabbroic intrusion, they were subsequently annealed due to the high regional heat flow. Pyrite cubes reached very large sizes and triple junctions developed indicating attainment of complete equilibrium. A second metamorphic overprint caused the development of epidote-amphibolite and later greenschist facies mineralogy in the silicate rocks. The presence of water led to re-mobilization of minor amounts of the ore body and its subsequent precipitation as a hydrothermal deposit.

A low temperature deformation episode was then imposed on the rocks. This was a very localized event at lower greenschist grade and gneissic textures were developed in the ores. These features are best developed in the north of the area where a late brittle shear passes over and through the ore body, at the mine. This shear is not present to the south where deformational features become less intense. Ores, within gabbroic rocks to the south of the mine, show little or no deformation due to the remoteness from the shear and also to the shielding effect that the gabbro had on any directed stresses. Post-tectonic recrystallization was very restricted and occurs only in the most highly strained parts of the ore where the stored energy was high enough for new grain development. The fact that many of the deformational features still remain, indicates that the deformation was at low metamorphic grade with no heat available for subsequent annealing.

CHAPTER SIX

STRUCTURE, LITHOLOGIES AND AGE RELATIONS AT EIKER

6.1 INTRODUCTION

The rocks of Eiker may be divided into three main lithological groups; the amphibolites, the acidic supracrustals and the sulphide ores. Northern and southern sub-areas can be distinguished (Fig. 6.1); both contain acidic supracrustals and basic rocks, but in the north the latter form intrusive stocks, while in the south they are concordant sheets within the supracrustals. From field evidence alone, it is not possible to tell whether these concordant basic sheets are part of the supracrustal sequence or later intrusions. The ore body is located entirely within a fault zone which trends NE-SW, roughly parallel to a major junction between basic intrusions and supracrustals (Fig. 6.1). The northern and southern sub-areas will be considered separately.

6.2 THE NORTHERN SUB-AREA

This constitutes the larger part of the study area and the divisions and relationships between the acidic and basic rocks are clear, in contrast to the southern sub-area. The sulphide-bearing shear zone is located entirely within the northern sub-area.

The supracrustals are leucocratic, massive to well-foliated rocks which form the lower lying land of the northern sub-area. Where developed, the foliation is fully penetrative and strikes NE-SW to the north of the shear zone, changing to E-W to the south of it (Fig. 6.2). This foliation is even constant across intrusive margins which are at a high angle to it, a feature taken to indicate that a high grade, metamorphic tectonic event occurred after the emplacement of the proto-amphibolite. The foliations dip predominantly at 50° to 70° to the north or north-west, varying both along and across the strike. There is no consistent change in dip as basic rock junctions are approached.

The strike of the foliations can be related to open folds and flexures. The gneisses were bent and folded around the more massive amphibolite intrusions, so that the foliations are now concordant to the intrusive contacts. Later features such as kink-bands are occassionally present. This later deformation is also displayed within the shear zone (see Section 6.3), and on a microscopic scale, with kink bands developed in biotites and frequent grain boundary granulation in quartz and feldspars. There is little evidence of an earlier, higher grade shearing in the gneisses, but rare intrafolial shear folds are developed, plunging steeply to the east or north-east. On a microscopic scale, the development of augen gneisses and recrystallized cataclastite textures are evidence of a higher grade deformation, which can also be related to features in the shear zone.

The amphibolite bodies of the northern sub-area form rounded outcrops and have the appearance of small stocks or bosses which may connect at depth. These rocks are usually well foliated, with the fabrics striking parallel to the contacts and dipping from 60° north or north-west to vertical. The amphibolites often contain xenoliths of gneiss at their margins and also further in from the contact. These xenoliths may well be roof pendants or blocks which were stoped by ascending basic magma. The discordant nature of the amphibolite-gneiss contact is evident on a small scale (Fig. 6.3) with minor discordancies and apophyses.

The actual contact may be sharp or more complex, when it is related to a progressive intrusion of basic material. This produces an inter-banding of gneiss and amphibolite, the relative abundance of each rock type being related to the proximity of the main amphibolite bodies. Alternations may occur down to the millimetre scale. The bands have sharp margins and the foliations are totally concordant with those of surrounding rocks. Towards the intrusion, this interbanded rock type may grade into a more mixed gneiss-amphibolite rock. In this case, amphibolite is dominant, but zones of relict gneissic material are still visible. The injection of basic material can be seen on a microscopic scale and all foliations are often highly distorted, in proportion to the extent of injection and assimilation of the gneisses.

In some cases, mobilization of the acid gneisses has occurred and led to the development of numerous felsic ptygmas. These ptygmas are very similar to metamorphic quartzo-feldspathic segregations (which occur frequently in the amphibolites), but field relationships of the mixed rocks, which only occur near the junctions, serve to distinguish the two types. The field relationships in the northern sub-area thus present unequivocal evidence for the intrusive nature of the amphibolites.

6.3 THE SHEAR ZONE

This zone widens northwards from a 2cm broad shear, ten metres south of working no.1, within a large amphibolite body (Fig. 6.3).

Further south there is no evidence of dislocation. The zone develops in a NNE direction and is clearly identifiable for 600 metres to shaft 11, where it has widened to a maximum of 20 metres. Further north, drift covers the shear zone which can, however, be traced as a tract of low and boggy ground trending in a north-easterly direction (Fig. 6.4).

The shearing in this zone varies in intensity. Discrete shear planes are common and are separated by zones of more or less highly-fractured rock. The shear zones are now composed of chlorite-actinolite schists, reflecting late brittle dislocations which were the most intense movements. They also bifurcate and die out along their strike, although even when they are not present, the fracturing in the zone is still more intense than in the surrounding country rocks. More competent units within the shear zone show necking and thinning, producing boudins (Fig. 6.5). The zone and individual shears within it, vary in both dip and strike, with dips ranging from 35° to 65° north-west. The upper margin of the shear zone often dips at shallower angles than the lower margin (i.e. the zone widens at depth, where the majority of the ore has been mined). Slickensides are frequently developed down the dip of the shear zone. The slickensides and boudinage freatures indicate that the sense of movement was essentially down dip, although it is not clear whether the shearing was in a normal or reverse sense. (The boudinage would suggest a tensional normal faulting.) The shear zone termination in the south-west suggests that the faulting may have been a scissor movement with its hinge at that end.

The shear zone is located near the junction of the supracrustal

FIG. 6.4. View north-eastwards from shaft 11 showing drift covered extension of shear zone.

FIG. 6.5. Shear zone with competent band (above hammer) showing necking and thinning. The band is bordered by two brittle shear planes.

gneisses and the amphibolite body. The southernmost part of the shear is located within the amphibolite, but rapidly bends towards the junction as the shear is followed north, whence it is located variously in both rock types and the interbanded units. To the north of shaft 8 (Fig. 6.3), the amphibolite body becomes much thinner in outcrop (10m) and forms part of an interbanded unit (Fig. 6.1). The thinning continues north until the basic rock is only one to two metres wide at shaft 11. Thus the amphibolite units within the interbanded rocks are intimately associated with the massive basic bodies and are subconcordant sheets or more irregular apophyses which have intruded the supracrustals at the margins of the stocks.

Two tectonically included pods of rock (in shafts 10 and 11) bear no mineralogical resemblance to any other rocks in the area. Both pods have brittle fractures which are continuous with the chlorite-actinolite shears outside them, indicating the presence of the pods during the late brittle deformation episode. The pod in shaft 11 is an ultrabasic body with a metamorphically altered sheath which contains relcs of the ultrabasic parent (Fig. 6.6 and 6.7). The pod has concentric ring fractures developed as a result of refraction of the brittle shears as they passed into the different rock type. The dip of these shears shallows towards the base and the top of the pod. The pod in shaft 11 now consists of muscovite, chlorite and calcite. The mineralogy and origins of the pods are discussed in Chapter Seven.

FIG. 6.6. Uncleaved ultrabasic pod (above hammer) and metamorphic rim (beneath hammer). (Shaft 11)

FIG. 6.7. Relic of ultrabasic pod (in centre) within the metamorphic rim.

6.4 THE ORE DEPOSIT

Massive and disseminated sulphide ores (predominantly pyrite with minor chalcopyrite, sphalerite and pyrrhotite) are restricted to the shear zone. Outcrops of the ore at the surface are very limited and, where present, intensive weathering has oxidised both ore and country rock producing a rusty coloured lithology which is usually very friable (Fig. 6.10).

The sulphides present at the surface show a high degree of deformation and are intimately associated with silicate rock fragments or individual silicate grains. Both amphibolite and acidic gneiss occur within massive sulphide and become impregnated by disseminated ore material. The silicate rocks are usually well-rounded fragments, commonly spheres and ovoids typical of the 'durchbewegung' fabrics (Figs. 6.8 and 6.9). Vokes (1970) explains that the texture described by this German word is produced by strong directional movement under higher grade metamorphic conditions. He states; "The movement producing such rounded forms (silicate rock) within the sulphide mass must be in the form of a thorough 'kneading' or 'milling' in which internal rotational movements dominate". These silicate spheres and ovoids are probably gangue and wall material, plucked away during shearing. Later recrystallization of both ore and silicates in the 'durchbewegung' texture has produced coarse grained rocks. The ores display many deformational features from brittle shear planes to recrystallized mylonite textures (see Chapter Eight).

The distribution of both massive and disseminated ore is random along both strike and dip, partly reflecting the extent of deformation that has taken place in the shear zone. The massive ore lenses out in

FIG. 6.8. Large rolled gneiss fragments in weathered sulphide host, forming a larger scale 'durchbewegung' fabric.

FIG. 6.9. Smaller scale 'durchbewegung' fabric of amphibolite in pyrite.

all directions and is often cut out by intersecting shear planes (Fig. 6.11). In shaft 11, the ore body is adjacent to a vein of black quartz two metres thick, but the junction is sharp and the quartz shows no signs of deformation while the adjacent sulphides have typical 'durchbewegung' fabrics. The quartz vein is a later stage mineralization (possibly Permian) and not associated with the sulphides. The massive ore is largely concordant with the general trend of foliations of the immediately adjacent country rocks, but does show small scale and low angle discordancies with amphibolite and gneiss. Despite this, the ore is virtually stratabound on a large scale.

6.5 THE LATE STAGE BASIC INTRUSIVES

Two small amphibolite dykes are present just to the south of shaft 8, striking N-S (Fig. 6.3). These dykes are unfoliated but display extensive brittle shearing. A very small outcrop of metagabbro (Fig. 6.1) has also intruded the supracrustals. These two rock types are not associated with the ore deposit, and are obviously later than the large amphibolite bodies. They may be similar to the late stage ('Vinor') amphibolite dykes at Grøsli.

6.6 THE SOUTHERN SUB-AREA

In contrast to the northern sub-area, the rocks of this sub-area are all developed in parallel bands, and there is often a gradation from acidic gneiss through hornblende gneiss and felsic-amphibolite to true amphibolite. This kind of gradation occurs both along and

FIG. 6.10. Rusty sulphide band between two gneiss layers.

FIG. 6.11. Rusty sulphide band (behind hammer) cut out by two intersecting shear planes with gneiss above.

across the strike. No intrusive junctions were observed on any scale. All of the units are foliated, striking NE-SW of E-W and generally dipping north at 60° to 70° (Fig. 6.2).

The origin of these basic rocks is dealt with in Chapters Nine and Eleven, but from field evidence alone, no definite conclusions can be drawn. In places, occurrence of interbanded amphibolite and gneiss, identical to those in the northern sub-area, suggest that the amphibolites may have been intrusives. Discontinuous bands of gneiss in the basic rocks can also be interpreted as xenolithic inclusions. However, there is a lack of even small scale discordancies which might be expected if the rock were intrusive. The presence of transitional mineralogies such as hornblende bearing gneiss and more felsic amphibolites could also be due to a supracrustal (sedimentary or volcanic) facies change. These mineralogies may also be explained by the mixing of basic volcanics and more acidic sediments. There is little or no evidence of shearing in the southern sub-area, unlike the sub-area to the north.

6.7 AGE RELATIONSHIPS

The Eiker area is part of the Kongsberg Series, which has always been considered as an integral tectonic unit. The rocks can thus be compared with those in the Grøsli area. Although no radiometric dates are available in the immediate vicinity of Eiker, the supracrustal gneisses and the later intrusive basic bodies are generally similar to those of the Grøsli mine area.

The supracrustals may thus be considered pre-Svecofennian in age. Samples of quartzo-feldspathic gneisses of the supracrustal succession

9km north-east of the Eiker area gave Rb-Sr isochron ages (for their metamorphism) of 1580 ± 50 m.y.B.P. (Jacobsen and Heier, 1978).

The intrusive basic rocks at Eiker are somewhat different from those at Grøsli, in that they are thoroughly metamorphosed and no relics of their original mineralogy remain. There are two possible explanations for the differences. Firstly, the basic rocks at Grøsli and Eiker may be of the same age, with differences solely due to a more thorough metamorphism in the Eiker region. A second possibility is that the intrusives at Eiker are earlier than those at Grøsli. This idea is supported by the presence of the late metagabbro intrusion at Eiker, which is identical to the larger bodies at Grøsli. The difference between the two intrusive groups could then be explained in terms of the greater age of the Eiker bodies.

The Eiker amphibolites may thus represent early 'Vinor' intrusive phases or even earlier injections associated with older dioritic gneisses and amphibolites (see Chapter One). Jacobsen and Heier (1978) have dated the intrusion of dioritic gneisses from localities 12km north of Eiker at 1520 ± 50 m.y.B.P. (i.e. Svecofennian in age). The age of the ore body is discussed in the following chapters.

CHAPTER SEVEN

PETROLOGY OF THE SILICATE ROCKS AT EIKER

7.1 THE AMPHIBOLITES

In the area as a whole, the amphibolites have varying modal proportions of amphiboles and grade to amphibole bearing gneiss. The amphibolite bodies in the northern sub-area have rounded intrusive forms and are the most basic, with common hornblende as the main amphibole (Table 7.1). The hornblendes form elongate prisms up to 5mm long, in the foliation, which is usually developed in these rocks. These hornblendes display a colourless to deep-green pleochroism, without the development of the blue-green colours present in many of the Grøsli amphiboles. Grain boundaries are usually straight although ragged ends to the crystals are common, while twinned and poikiloblastic forms occasionally occur. Rare monomineralic bands of hornblende (up to 2cm thick) are present and are composed of elongate crystals up to 2mm long which are often interlocked in a tesselate pattern. Associated with these bands are layers of tremolite which have a non-directional fabric and display an equigranular tesselate texture. A small amount of calcite is usually associated with these bands, which seem to have formed by Ca metasomatism.

Plagioclase feldspar is the second essential constituent of the amphibolites. It forms stubby laths up to 0.7mm long but more often very irregular forms are developed. Twinning is on albite, pericline and Carlsbad laws, although some untwinned crystals are also present. Normal zoning is common, with rarer reverse zoning, giving oligoclase-

TABLE 7.1. ELECTRON MICROPROBE ANALYSES OF AMPHIBOLES IN THE EIKER AREA

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	40.62	40.13	40.41	40.54	39.89	52.49	52.10	50.24	48.54	52.54	52.95	54.20
TiO_2	1.13	1.00	1.13	1.03	1.03	0.04	0.12	0.15	0.43	0.20	0.10	0.13
Al_2O_3	15.31	14.07	14.94	14.23	14.74	3.95	4.98	5.34	12.20	5.11	5.60	4.74
FeO	19.75	20.50	20.41	20.51	20.40	7.00	6.39	11.88	7.18	5.10	5.44	4.92
MnO	0.26	0.30	0.26	0.36	0.48	0.36	0.48	0.46	0.53	0.54	0.54	0.43
MgO	7.53	7.34	7.29	7.10	7.27	19.74	19.03	15.72	15.61	19.32	19.72	20.13
CaO	11.39	10.93	11.41	11.19	11.04	12.28	12.66	11.56	11.51	12.54	12.71	11.56
Na_2O	2.42	1.90	2.11	1.78	1.67	1.31	1.52	1.69	1.73	1.47	1.68	1.22
K_2O	0.84	0.86	0.90	0.86	0.80	0.11	0.22	0.46	0.48	0.14	0.22	0.15
TOTAL	99.25	97.03	98.86	97.60	97.32	97.28	97.50	97.50	98.21	96.96	98.96	97.48

FeO = total iron as FeO

Analysis 1 to 5. Hornblendes from amphibolites of the northern sub-area.

Analysis 6 to 9. Hornblendes from rocks with 'durchbewegung' fabrics in shear zone.

Analysis 10 to 12. Actinolites from rocks with 'durchbewegung' fabrics in shear zone.

andesine compositions (An_{22-46}). Post-crystallization deformation of the multiple twins is a common feature. Quartz is present as an essential, but minor constituent and forms individual crystals up to 0.5mm across or aggregates of smaller irregular grains displaying undulose extinction. The mineral also occurs as later vein quartz, typically with crystals with highly irregular boundaries. Garnet is common and forms sub-idioblastic crystals up to 0.8mm across. These amphibolites are thus the products of total recrystallization of medium grade amphibolite facies metamorphism (as described by Winkler, 1978).

In the south of the Eiker area, the amphibolites are interbanded with the gneiss. They contain up to 50 percent modal amphibole (in contrast to the stocks of the northern sub-area which contain up to 70 percent), and grade into hornblende gneisses with 20 percent or less modal amphibole; the other major constituents are plagioclase (An_{28-40}), biotite, garnet and quartz. Accessories in these rocks include sphene, epidote and chlorite. The amphibole of these rocks is common hornblende, optically identical to the amphibole in the more basic amphibolites of the northern sub-area but no alteration to tremolite is present.

At the margin of the northern intrusive amphibolite stocks the injection of basic material has given a marginal zone of mixed gneisses which are hybridized to varying degrees with clots, stringers or single crystals of basic material in the gneisses. The crystals are xenoblastic and no larger than 1mm in their longest dimension. Foliation is absent and the crystals take up an orientation in the direction of injection with the result that the rock appears highly contorted, displaying varying mineral orientations. The crystals are highly ragged in outline and often chloritized. The mixed gneisses are not in complete equilibrium, partly because of later incomplete retrogressions and partly due to the fact that the rock is a hybrid in which the invading basic material has not equilibrated with the acidic host minerals. There is a complete gradation from gneiss through mixed gneiss to amphibolite although the presence of garnet as a major constituent seems to indicate a gneiss precursor.

7.2 THE SUPRACRUSTALS

These rocks are variable acidic and intermediate gneisses and schists. All rocks contain quartz, plagioclase, microcline and biotite. In addition hornblende, garnet, muscovite, sillimanite and cordierite may form major or accessory constituents. Epidote and magnetite are common accessories. Quartz and feldspars generally form granoblastic textures of varying grain size, averaging 0.2 - 0.6mm but increasing to 2 - 3mm in some. Grain boundaries vary from straight (often with polygonal junctions) to more highly serrated forms. As the degree of complexity of grain junctions increases, variability of grain size and degree of straining shown by the grains also increases. Plagioclase, though only rarely a major component, is most abundant in the hornblende gneisses where it is of sodic andesine composition. In the more acidic rocks it is oligoclase. Alteration of the feldspars (away from the shear zone) is slight and restricted to sericitization at grain boundaries and along cracks (Fig. 7.1).

In some gneisses near the shear zone, cataclasis has produced augen gneisses with a larger than average variation in grain size and

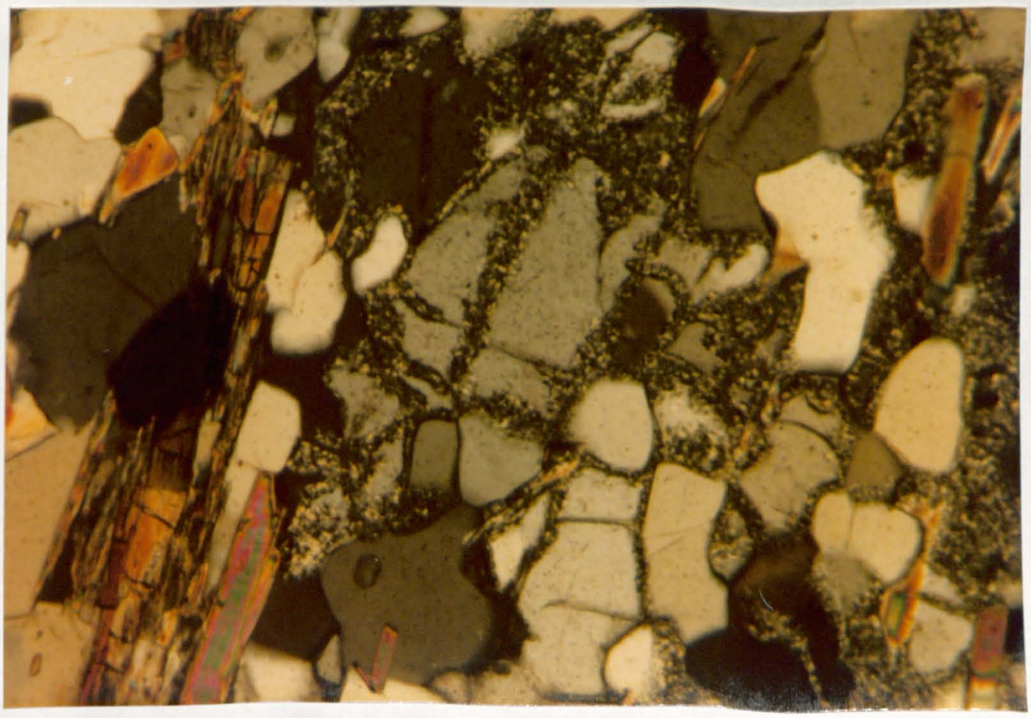


FIG. 7.1. Sericitization of feldspars along cracks and grain boundaries. XPL. X12.5.

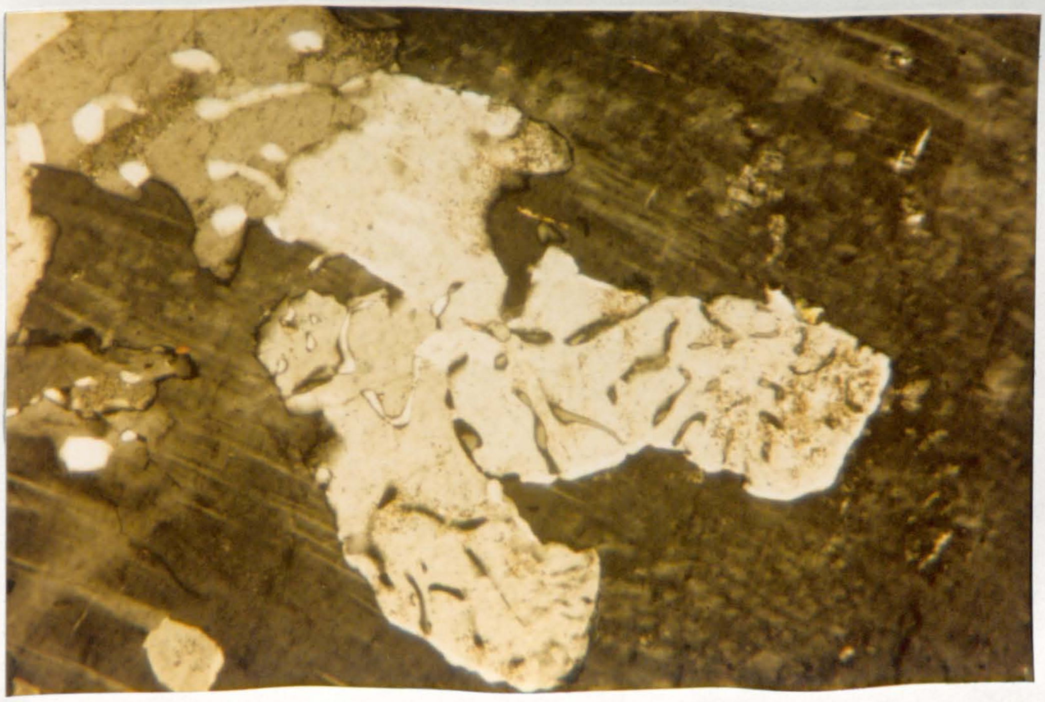


FIG. 7.2. Microcline perthite with inclusions of partially resorbed, myrmekitic plagioclase. XPL. X31.5.

highly irregular crystal margins. Large relict microcline porphyroblasts (up to 5mm across) are common and are highly fractured, sometimes with a mortar texture of small grains around their margins, developed by movements after the augen formation. Microcline perthite occurs with inclusions of partially resorbed, myrm kitic plagioclase (Fig. 7.2).

Biotite and muscovite (where present) have usually imparted a crude foliation and when in greater proportions have produced a more schistose fabric. They form elongate grains up to 3.5mm long and the biotite is present in both green and brown forms although the latter is more common. Biotite also occurs around grain margins and in the cracks of the quartz and feldspars. It shows evidence of tectonism with the development of kink bands (Fig. 7.3) and rotational fabrics, including spiralling (Fig. 7.4). The biotite crystals are usually ragged and show at least partial chloritization. Second generation, static growth grains are also present. Hornblende often occurs as prisms up to 3mm long or as poikiloblastic, equigranular crystals, often intergrown with biotite. Hornblende and sillimanite are mutually exclusive (a function of rock chemistry). Biotite-hornblende mats are often folded around garnets which may be up to 30mm across and of two generations. The earliest are pre-tectonic forms, now exhibiting rounded outlines, with matrix material folded about them. Inclusion trails within garnets are often discordant to the external fabric, indicating rotation during tectonism. The second generation garnets are highly irregular 'sieve' forms which have grown across the matrix foliations and are obviously due to static growth.

Sillimanite occurs as needles up to 5mm long, parallel to the foliation. Where present on the fault-planes, in the shear zone,

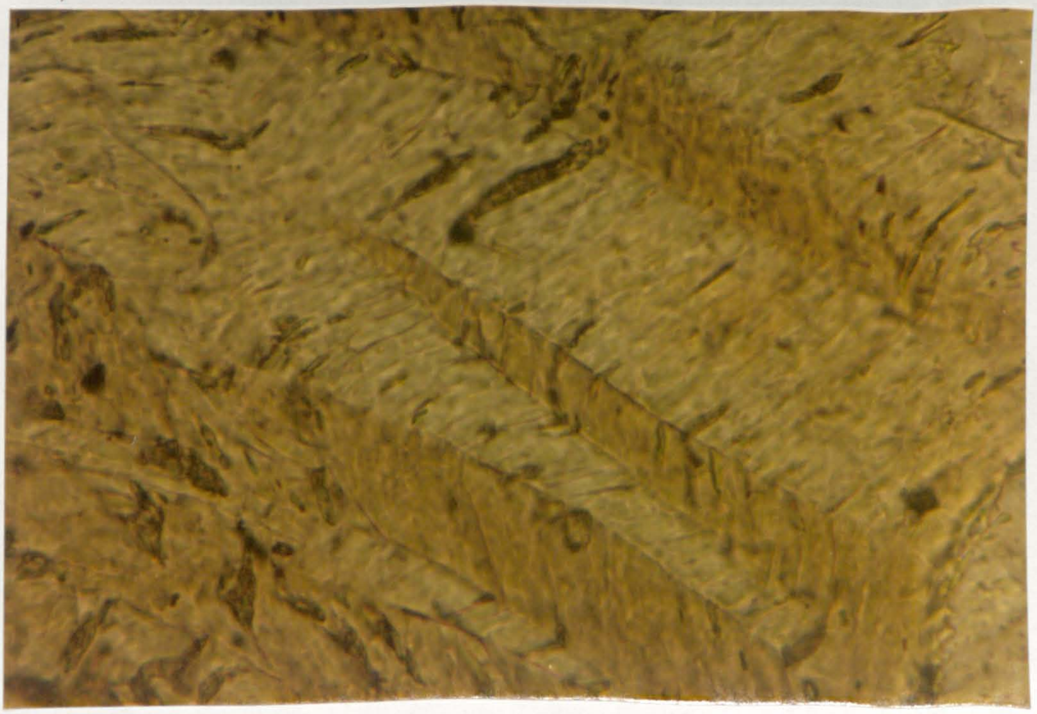


FIG. 7.3. Kink bands in biotite. PPL. X50.

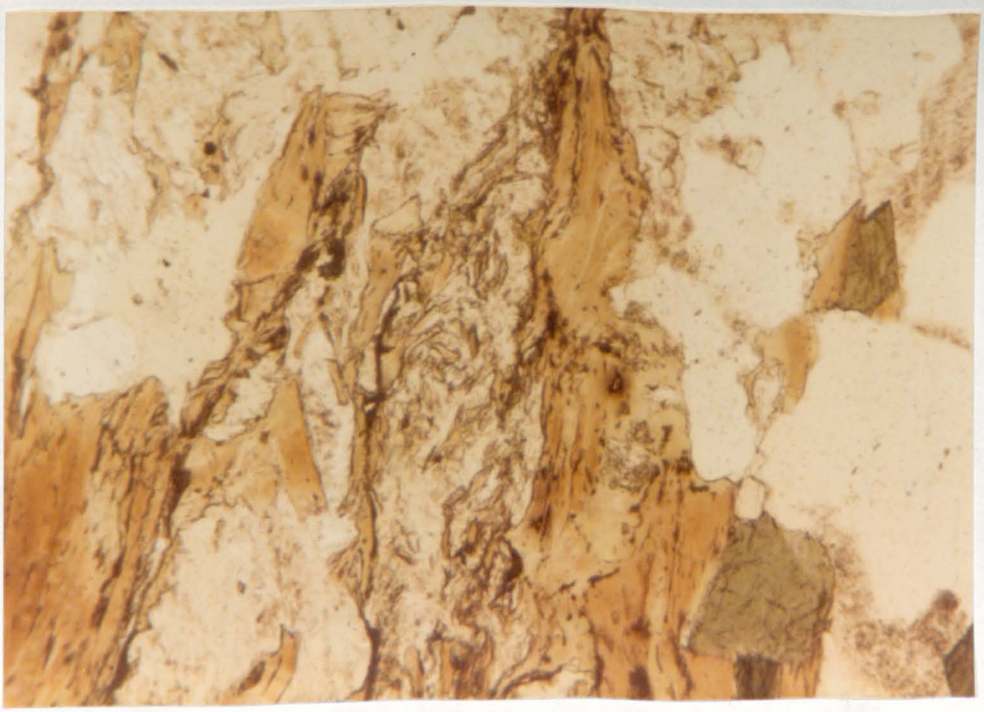


FIG. 7.4. Fractured and 'spiralled' biotite developed in augen gneiss. PPL. X12.5.

cracked sillimanites have often been re-orientated parallel to the slickenside directions. Where visible on fault planes, these are usually weathered (Fig. 7.5). Sillimanites away from the shear usually occur enclosed within second generation garnets, but folded around those of the first generation.

7.3 THE SULPHIDE BEARING ROCKS

Major developments of sulphides (other than accessory amounts), are limited to those rocks within the shear zone. In the field, the ore appears to be strongly associated with melanocratic rocks, but microscopic examination reveals that much acidic gneiss material is also present but has very often been darkened by chloritization. Interpretation of sulphide-silicate mineral relationships is difficult since later alteration has, in most cases, completely replaced earlier minerals. However, occasionally less altered mineralogies are preserved within the shear zone; as rolled silicate masses (within 'durchbewegung' fabrics) and between the main shears, as lenses. Most of these relict rocks are similar to those (described above) which surround the shear zone. However, some of the basic rocks are composed of variable mixtures of chloritized actinolite and rarer hornblende, biotite, gahnite (Fig. 7.6 and Table 7.2) and sulphide.

The rolled silicate masses range in size from 3 or 4mm to several tens of mm and represent gangue and country rock torn off the walls as the ore body was sheared. These ovoid bodies occasionally exhibit a relict high-grade equilibrium texture with straight grain boundaries, but usually the fabric shows evidence of much greater stress. Individual

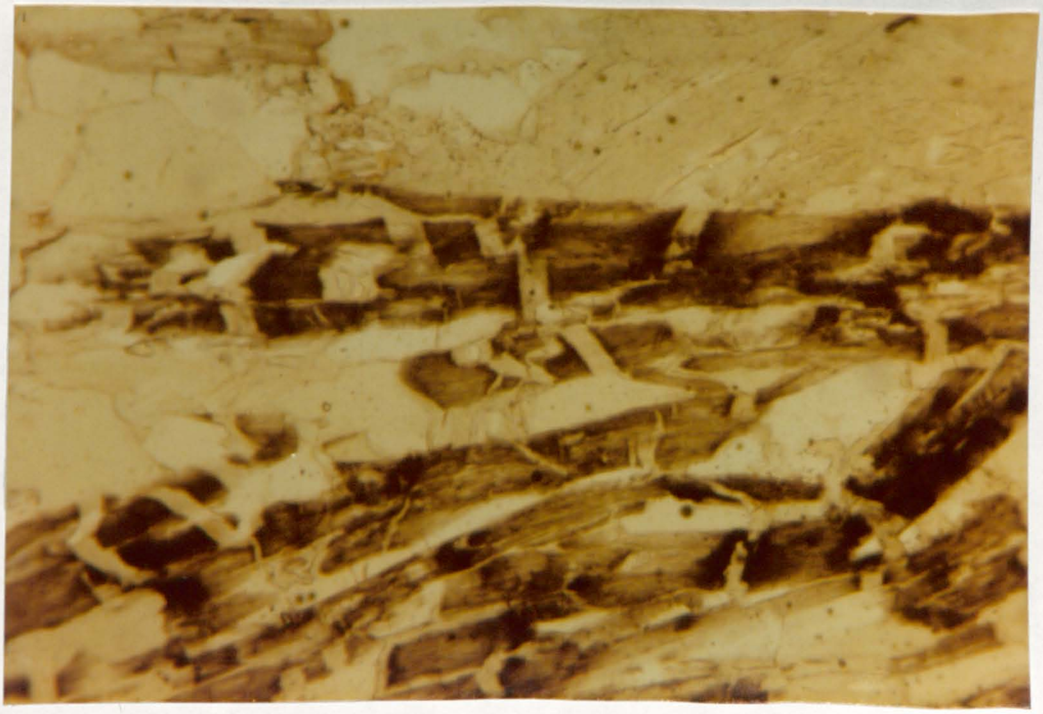


FIG. 7.5. Weathered, cracked sillimanite crystals on surface of fault plane. PPL. X31.5.

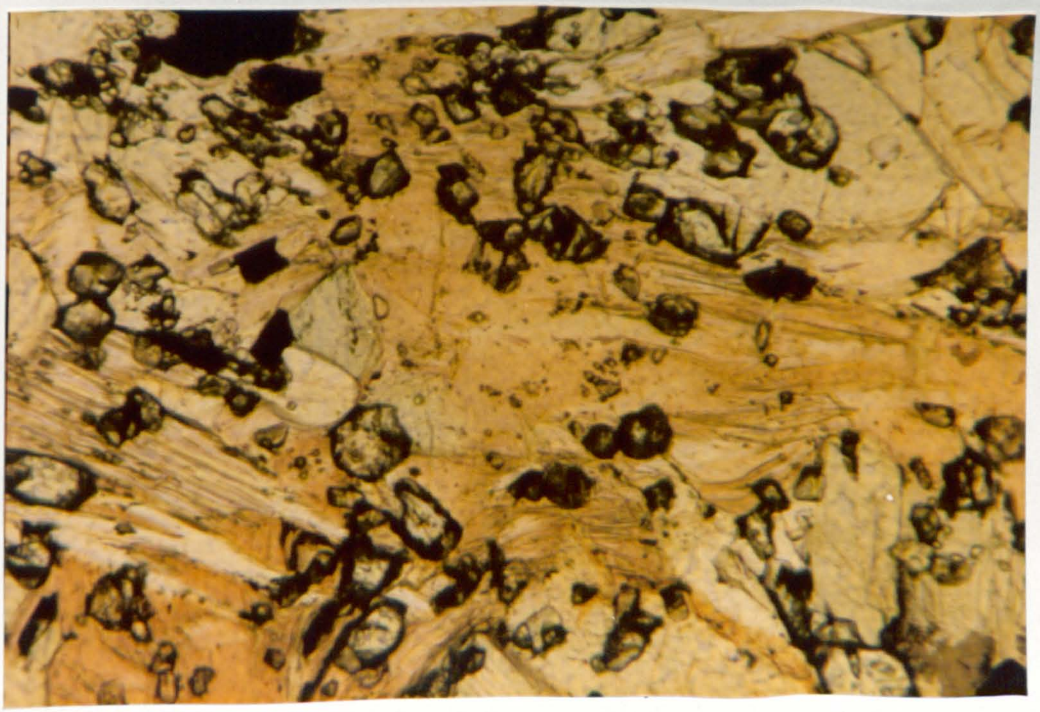


FIG. 7.6. Light green gahnite crystals with biotite, amphibole and chlorite in rolled silicate, from 'durchbewegung' rock in shear zone. PPL. X12.5.

TABLE 7.2

ELECTRON MICROPROBE MINERAL ANALYSES

	1	2	3	4	5
SiO_2	37.29	38.64	32.26	25.24	0.36
TiO_2	0.00	0.03	0.03	0.00	0.00
Al_2O_3	21.69	24.54	18.05	16.82	59.50
FeO	30.06	12.15	24.84	36.55	7.07
MnO	0.74	0.35	0.20	0.23	0.39
MgO	7.54	0.09	11.82	8.26	3.54
CaO	2.21	23.44	0.00	0.00	0.00
Na_2O	0.00	0.00	0.64	0.17	0.00
K_2O	0.00	0.00	0.16	0.00	0.00
ZnO	N.D.	N.D.	N.D.	N.D.	28.92
TOTAL	99.53	99.21	88.00	87.27	99.78

FeO = total iron as FeO

N.D. = not determined

Analysis 1. Garnet from acidic supracrustal gneiss.

Analysis 2. Garnet from sulphide-chlorite-calcite assemblage in shear zone.

Analysis 3. Chlorite from sulphide-chlorite-calcite assemblage in shear zone.

Analysis 4. Chlorite from reaction rim between sulphide and amphibole (see Fig. 7.9).

Analysis 5. Gahnite from sulphide-chlorite-calcite assemblage in shear zone.

prisms of amphible up to 3mm long are present, but the matrix is largely composed of xenomorphic grains 0.2 - 0.7mm across, tightly interlocked with sutured and irregular boundaries. Most of the constituent grains have strong undulose extinction and the biotites display very well developed kink bands.

The amphiboles associated with the ore, where not extensively chloritized, are of actinolitic character (Table 7.1). However, very rarely, their cores are of a more aluminium rich hornblende. These hornblendes are of somewhat different composition to those in the major amphibolite bodies (Fig. 7.7) and it is suggested that they represent recrystallized gangue associated with the ore. The actinolitic amphiboles developed during the brittle greenschist movements and recrystallized from both the gangue and the major amphibolite bodies. A later chloritization then affected the shear zone and other original material has been totally recrystallized, either to biotite-actinolite or chlorite-carbonate assemblages which (together with sulphide) now form the vast majority of the rock compositions within the shear zone. Chlorite is by far the most common silicate mineral in the 'durchbewegung' fabrics, while grossular garnet (Table 7.2) is also occasionally present.

7.4 THE TECTONICALLY INCLUDED PODS

Within the shear zone, in shafts 10 and 11, are tectonically included pods. In shaft 11 the xenolith is an ultrabasic rock and its metamorphic envelope. A core of serpentinite can be recognized in which relict, highly altered olivines (up to 1mm long) are present. The crystals have usually developed a green body colour typical of

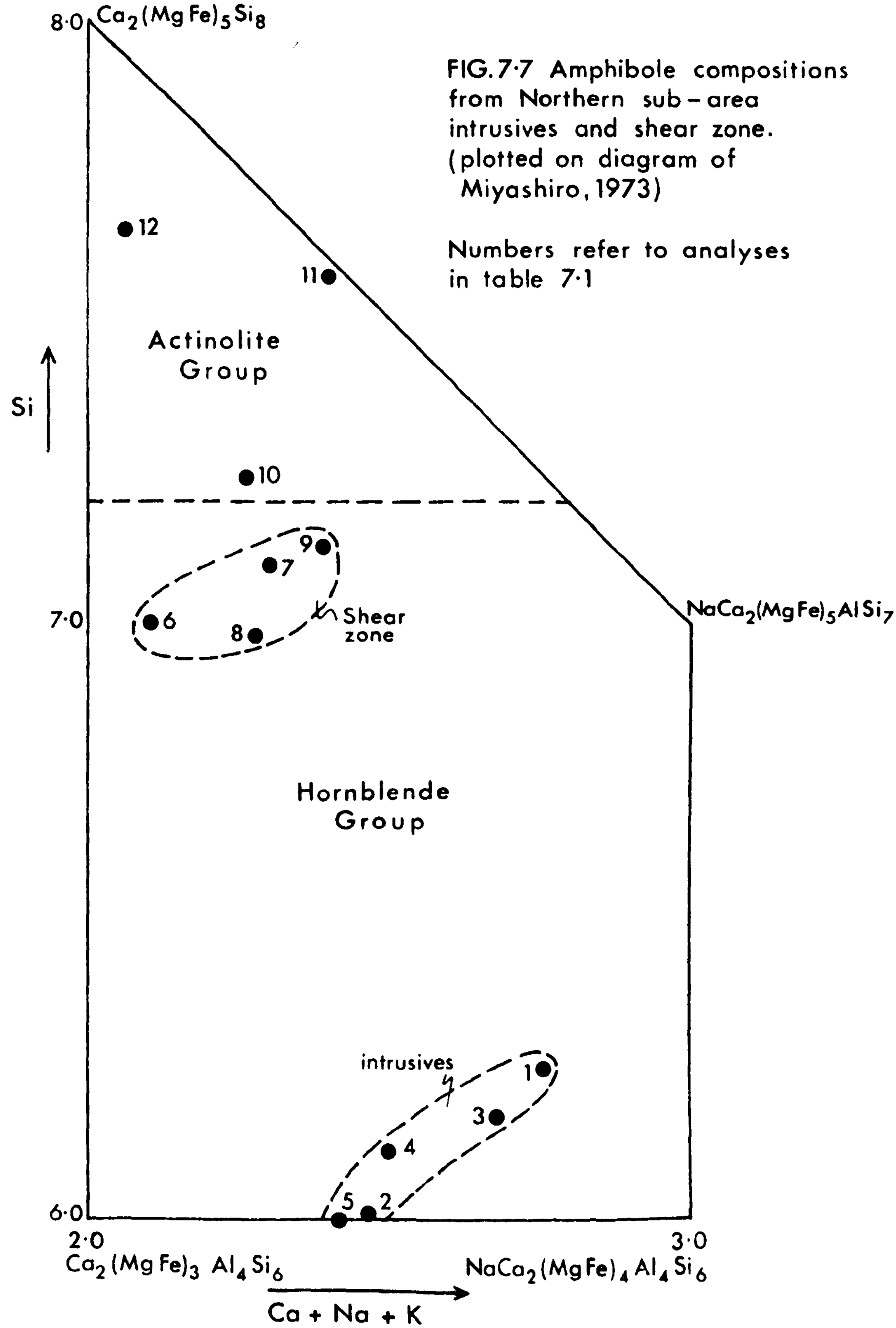


FIG. 7·7 Amphibole compositions from Northern sub-area intrusives and shear zone. (plotted on diagram of Miyashiro, 1973)

Numbers refer to analyses in table 7·1

alteration to bowlingite; they are dusted with magnetite and cut by brucite veins. The groundmass is totally serpentinized and composed of mats of minute crystals. The rock also contains large (up to 3mm long) crystals of static chlorite and occasional aligned chlorites in shears. Sulphides, picked-up during the emplacement of the pods, are highly xenomorphic, cut by serpentinite veins and exhibit signs of re-mobilization on a limited scale with veinlets cross-cutting the fabric.

Around the serpentinite body (and containing relics of it) is a metamorphic sheath of tremolite and calcite. Its formation from a serpentinite or peridotite pre-cursor would require an influx of calcium. This rock provides further evidence for movement of calcium-rich metasomatizing fluids in the Eiker shear zone. Accessories in both sheath and core rock include chlorite, and later replacive calcite which occurs in two forms; highly irregular, turbid patches are associated with re-mobilized sulphide which has dusted them and, in places, the sulphide has coalesced into discrete grains with the calcite losing its turbidity and forming good crystals.

A second tectonically included body (in shaft 10) appears to have had a different original composition and has now been converted to a muscovite-chlorite schist with more highly altered patches (up to 5mm across). These patches are the remnants of garnets, often retaining good idioblastic outlines. The patches are very turbid and consist of an aggregate of opaques with chlorite and muscovite ($\pm$ epidote). This pod appears to have been a supracrustal rock which was emplaced in the same manner as the ultrabasic body. Both bodies are so large (6 to 12m) that they are surrounded by both sheared ore and silicate rock.

7.5 THE LATE BASIC INTRUSIVES (METAGABBROS AND AMPHIBOLITES)

These occur in two dykes and one small intrusive stock (all located in the northern sub-area). They have undergone amphibolite grade metamorphism, but are non-foliated and have suffered no shearing except the late-stage, brittle movements. The dykes represent metamorphosed dolerites and have developed a very regular granoblastic texture with an average grain size of 0.1 - 0.2mm. Hornblende and plagioclase together form 95% of the rock with accessory quartz, magnetite and sphene. The small intrusive stock has a less equilibrated mineralogy; it is a typical amphibolitized gabbroic rock (identical to those in the Grøsli region), exhibiting a good ophitic texture with relict pyroxenes and two amphibole pseudomorphs of pyroxene (giving a rim and core). Epidotization is absent and the plagioclases are fresh, having a composition of An_{32} to An_{70} in the zoned specimens.

7.6 METASOMATISM AND FLUID MOVEMENT

Mobility of fluids in the Eiker area has not caused great alteration, except in the shear zone. Here, the intense shearing of interbanded amphibolites and gneisses has combined with the alteration to form a complex zone. The shearing has broken and displaced rock units, making the tracing of individual bands difficult: it has also assisted alteration by fluids, which is complete in many cases.

A sequence of progressive alteration can be determined, commencing with chloritization and sericitization of the mafic minerals and feldspars. Amphiboles are progressively altered along their cleavages and inwards from crystal margins. Feldspar grains are sericitized, becoming densely clouded with a mat of pinnate micas, while quartz grains are left unaltered. Biotite laths first lose their body colour and then alter to chlorite. The next stage is a coalescing of these altered patches, so that large areas of the rock are composed of a sericite-chlorite mat with relict quartz inclusions. Often, recrystallization has produced coarse grained rocks, with crystals up to 1cm long. As alteration proceeds, calcite forms a more major component and the result is often a chlorite-calcite-sulphide ($\pm$ quartz) rock bearing no resemblance to the protolith. Two stages of alteration are present in the shear zone; firstly an early tremolite-forming, metasomatic event and secondly an alteration comprising chloritization, followed by calcite development. The second stage has caused by far the greatest extent of alteration, the effects of which have been described above.

The early metasomatic event is reflected by grossular garnet and gahnite in the shear zone, the tremolite sheath on the serpentinite pod and the tremolite in the amphibolites in the northern sub-area. The gahnite (Zn-rich spinel) provides evidence that fluids have passed through the ore body. Later chloritization produced chlorite veins cross-cutting the fabric and the sulphides (Fig. 7.8) as replacements or infillings of brittle cracks caused by shearing in the ore. In some cases the individual sulphide fragments are highly irregular in shape, but still retain a cubic outline suggesting replacement of the ore in situ. In other cases, the individual fragments have several straight edges but a totally irregular aggregate outline, indicating a tectonic rather than replacive fragmentation.

Chlorite occurs as static growth crystals up to 1cm across. Where this texture is developed relict properties of the replaced minerals are

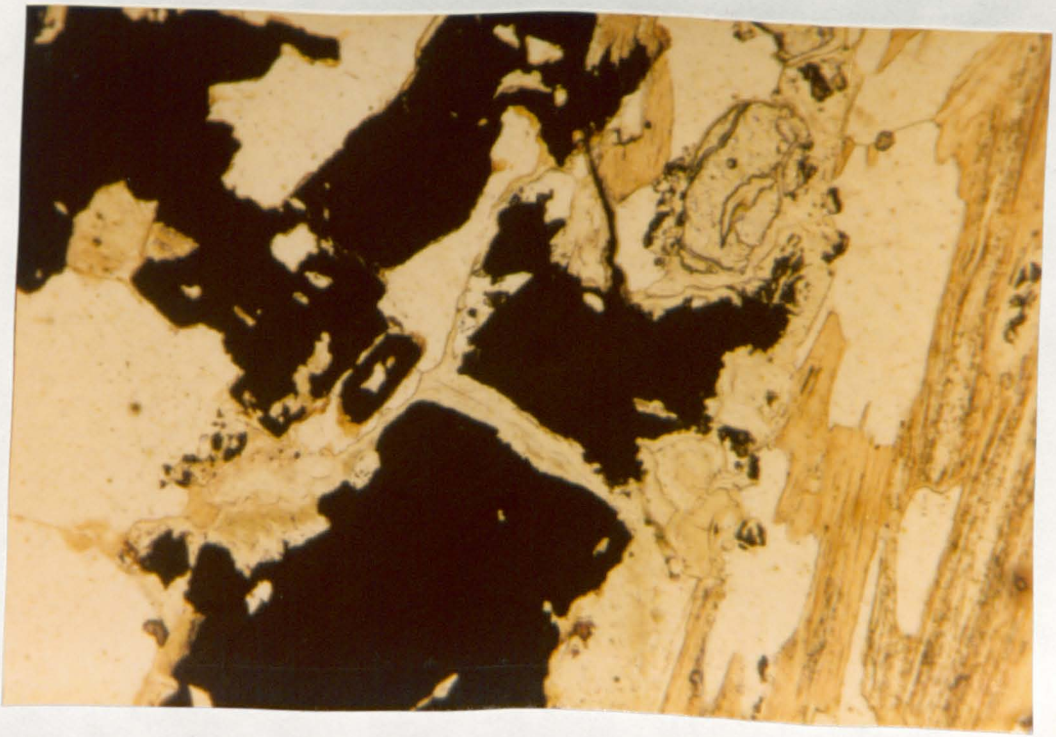


FIG. 7.8. Chlorite veins cutting sulphides. PPL. X12.5.

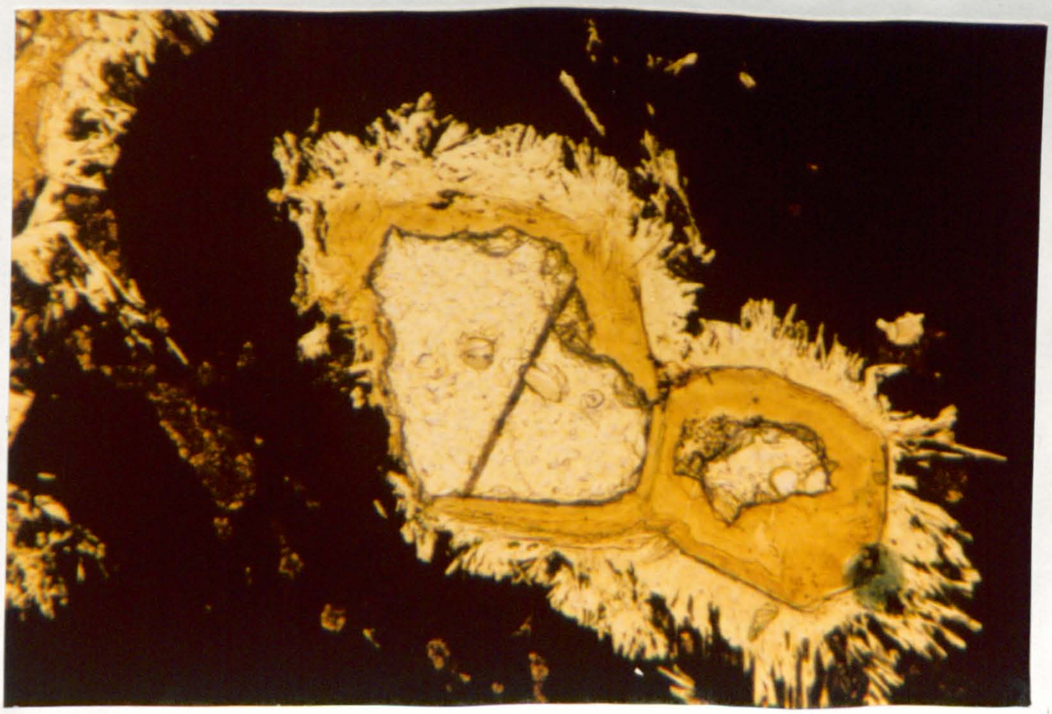


FIG. 7.9. Fe-rich chlorite developed from reaction between sulphide and amphibole (brown). PPL. X31.5.

still visible. These include twins and 120^{o} cleavages of amphiboles. Thus the late stage brittle shearing which effected the chloritization was not totally pervasive and, between the shear zones, static chloritization occurred. Within the shear zones, chloritic schists have typically developed. Mats of chlorite crystals (0.3 - 0.5mm long) have totally replaced the rock and sulphides present have been drawn out into trails of fragments (Fig. 7.10). In many cases a chlorite reaction rim has developed between sulphides (which now show corrosion textures) and amphiboles (Fig. 7.9). In these cases, the sulphide is being replaced and iron is being derived from the ore for the formation of the chlorite which is iron-rich compared to chlorites formed from amphiboles etc. (Table 7.2). Rolled silicate material which has not been totally chloritized invariably has a sheath of chlorite. Garnets have an inner amphibole rim, which in turn has an outer rim of chlorite against sulphide.

Overprinted, onto the highly altered fabric in the shear zone, has been a later stage carbonate development. Calcite with minor quartz and alkali feldspar has veined and replaced all previous minerals, in many cases. The degree of alteration is linked directly with the degree of brittle fracture of the ores. The calcite is commonly in the form of highly irregular patches which are highly dusted with partially replaced opaque impurities of the groundmass. With development of better crystal forms, the turbidity decreases and rhombs of calcite up to 1mm across develop. In ore which has suffered strong fracturing, the cracks are infilled by calcite or calcite-quartz-alkali feldspar aggregates which have often developed a columnar habit perpendicular to the sulphide boundary (Fig. 7.11).

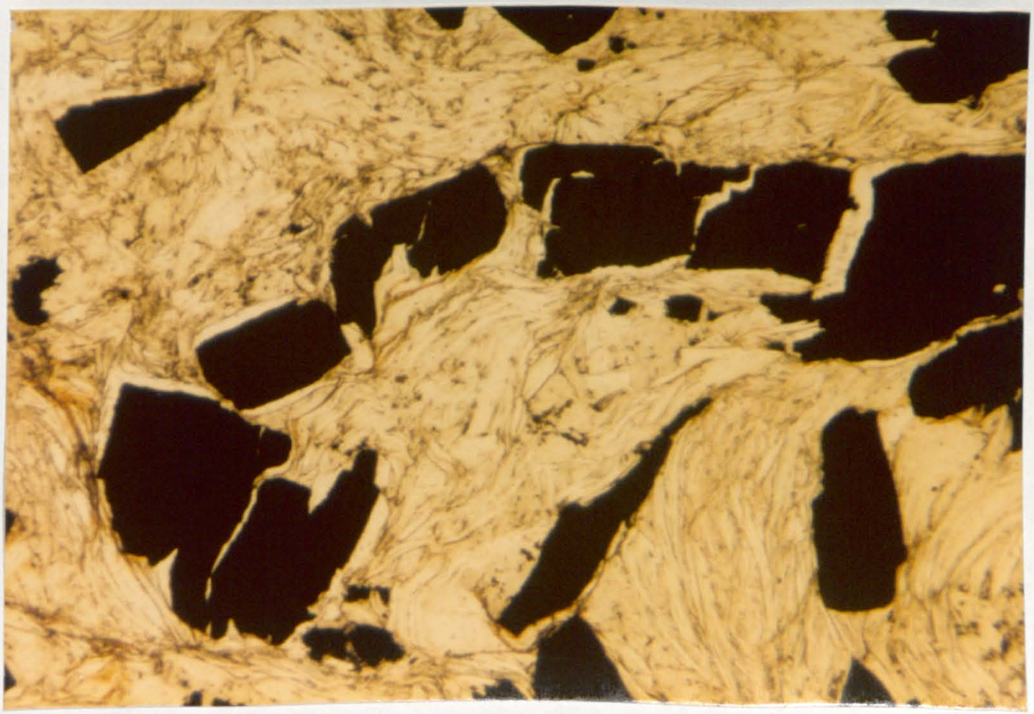


FIG. 7.10. Sulphide fragments in chloritic matrix of sheared rock. PPL. X31.5.

FIG. 7.11. Calcite veins cutting sulphide. PPL. X12.5.

Evidence of re-mobilization of the ore body is almost entirely absent and it is concluded that the extent of alteration is due primarily to the highly reactive nature of the rocks, brought about by their shearing, rather than by a high volume of fluid. This is borne out by the fact that alteration outside the shear zone is very slight. However, a small amount of re-mobilization has occurred. Rare veinlets (0.2 - 0.5mm wide) of sulphide associated with calcite, cut across chlorite-sulphide fabrics and occasionally coalesce into aggregates which enclose other minerals in a typical interstitial relationship. The sulphides have the same assemblages as the main ore body and, in addition, hydrated iron oxides. Another feature of the re-mobilized material is the iron-stained rim in the silicate host minerals around the ore vein (Fig. 7.12).

7.7 SUMMARY OF THE EFFECTS OF ALTERATION

The assemblage sulphide-chlorite-calcite-gahnite $\pm$ epidote is common in the ore zone. All of these gangue minerals suggest fluid activity during the brittle greenschist movements of the shear zone. The earliest metasomatism produced the minerals like tremolite, grossular and gahnite which were associated with the formation of the actinolitic amphiboles in the ore body. Later fluid activity caused extensive chlorite and carbonate development. These two minerals were produced in such abundance that chlorite-carbonate-ore assemblages are often observed. Therefore movement of fluids appears to have occurred throughout the brittle shearing episode. A summary of events is produced in Table 7.3.

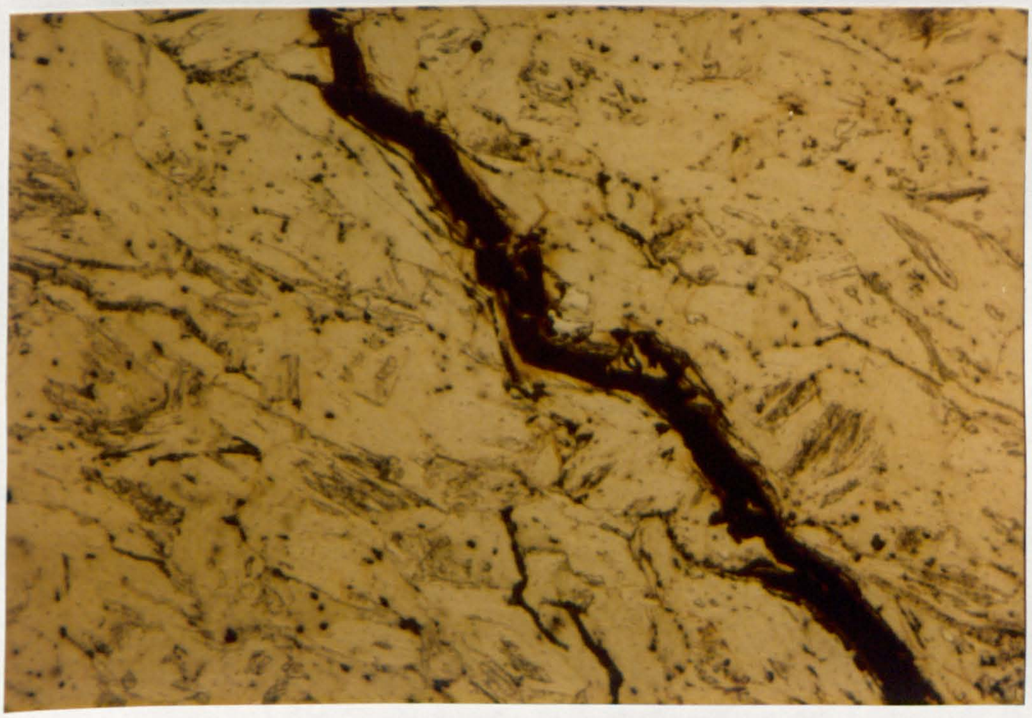


FIG. 7.12. Thin, re-mobilized sulphide veinlet, with staining of silicates on margin. PPL. X31.5.

TABLE 7.3

SUMMARY OF EVENTS IN THE SHEAR ZONE (WHICH LOCATES THE ORE BODY)

Intrusion of basic bodies	
Mid-amphibolite facies metamorphism with directed stress.	Sveconorwegian orogeny
Early shearing: formation of cataclastites, rolling of original garnets, production of 'durchbewegung' fabrics, inclusion of tectonic pods.	Mid-amphibolite facies
Short period of static recrystallization. Growth of later garnets. Introduction and amphibolitization of gabbro and dolerite dykes. Recrystallization in ore body.	
	decreasing metamorphic grade
Late shearing: (greenschist facies grade). Formation of actinolite-chlorite schists where movement occurred. Sequence of metasomatic and metamorphic retrogressive changes: Formation of calcium bearing minerals e.g. tremolite, grossular and gahnite. Recrystallization of amphiboles, (associated with ore) to actinolite-chlorite assemblages.	Greenschist facies
Later event of extensive chloritization which affected whole shear zone.	
Later carbonate alteration and very limited sulphide re-mobilization.	

7.8 POSSIBLE ORIGINS OF THE ORE BODY

The silicate-ore assemblage in the shear zone is the product of thorough recrystallization, together with metamorphic and metasomatic modification. The original ore-gangue associations have thus been almost totally obscured, making conclusions from petrographic studies difficult.

Three possible origins could be suggested for the ore. Firstly, it may have been part of the supracrustal sequence, either as a sedimentary or bedded volcanogenic ore. Secondly, it may have been magmatic, associated with the intrusive rocks. Indeed, many ore deposits do occur at the margin of intrusive basic bodies. Thirdly, the deposit may have been introduced and be unrelated to surrounding silicate rocks.

The possibility of the ore being volcanogenic is the most plausible explanation and is discussed in Chapter Eleven. At first sight, it seems possible that the ore may have originated from the intrusive amphibolite, with which it is often intimately associated. The objection against a primary basic magmatic origin is discussed in Chapter Eleven.

The ore body which has associated high grade 'durchbewegung' fabrics was obviously emplaced before the chlorite and calcite forming events. There is no evidence of substantial re-intrusion of ore material into the higher grade rocks such as the pyritic mylonites.

Thus, the ore body at Eiker may have two possible origins. It may represent a hydrothermal deposit but not associated with any of the late chloritic or calcitic material now present. It may represent part of a volcanogenic-sedimentary sequence of pre-Svecofennian age, but is unlikely to be associated with the basic intrusives. Whatever the true origin of the ore body, it was present prior to the commencement of the shearing in the area and thus probably represented a pre-existing structural discontinuity which located later movements.

CHAPTER EIGHT

THE ORE DEPOSIT AT EIKER AND ITS METAMORPHISM

8.1 INTRODUCTION

The mineralogy of the deposit is dominated by pyrite with subordinate pyrrhotite and minor sphalerite and chalcopyrite. Magnetite is present as an accessory, while microscopic flakes of gold are visible within the pyrite (Fig. 8.12). Gangue minerals include chlorite, muscovite-sericite, phlogopite-biotite, actinolitic amphiboles, quartz, calcite and the zinc rich spinel, gahnite.

The ore deposit has developed a gneissic fabric and is highly fractured and sheared. Shearing has been intense enough to produce mylonite bands in massive ores and rolled silicate bodies in sulphide matrices (i.e. 'durchbewegung' fabrics) are common (see Section 6.4). The ores underwent high grade regional metamorphism under directed stress with pre-, syn- and post-tectonic recrystallization features. A later phase of brittle shearing was followed by partial annealing of the softer sulphides.

The complex metamorphism is shown by the textures of the sulphides. These include rolling of pre-tectonic crystals, pressure shadow crystallization, syn-tectonic rotational growth (shown by inclusion trails), post-tectonic growth across directed fabrics and secondary overgrowths on earlier crystals.

8.2 MINERALOGY

Pyrite occurs in many habits which reflect the complex history of the ore body. It may occur as large porphyroblasts up to 2 cm across which are invariably cracked or broken. Some cubes are more highly fractured and granulated, the cube being sheared out into a trail of irregular fragments (of average size 0.1 - 0.2 cm across, Fig. 8.1). The line of these fragments can often be traced back to a remnant cube. Perfect cubes of about the same size as the granulated fragments are also common and may, or may not, be poikiloblastic. In some cases, the large porphyroblasts have been rolled to produce a rounded outline (Fig. 8.2). More rarely, spiral inclusion trails also occur in these types (Fig. 8.3). A second common form is that of a very irregular xenoblastic mass which may contain a dense dusting of minute inclusions or be completely clear. Small cubes may be present at the edges of such masses or occur enclosed within them. The origins of these various forms are discussed later.

Pyrrhotite is the mineral in which the gneissic fabric has been retained. Rolled silicate rocks, pyrites and sphalerites often have a pyrrhotite matrix deformed around them in much the same way as micas may deform around more resistant garnets (Fig. 8.4). The pyrrhotite is usually recrystallized to a mosaic of equidimensional crystals (Fig. 8.5) but often this has not destroyed the gneissic banding. Strain extinction and sub-grain development is common. Triple junctions are present but more often, higher energy boundaries exist between the grains. Often within the matrix, pyrrhotite has developed a mosaic of small interlocking crystals, whilst pyrite has formed idioblastic cubes which appear to have grown across the

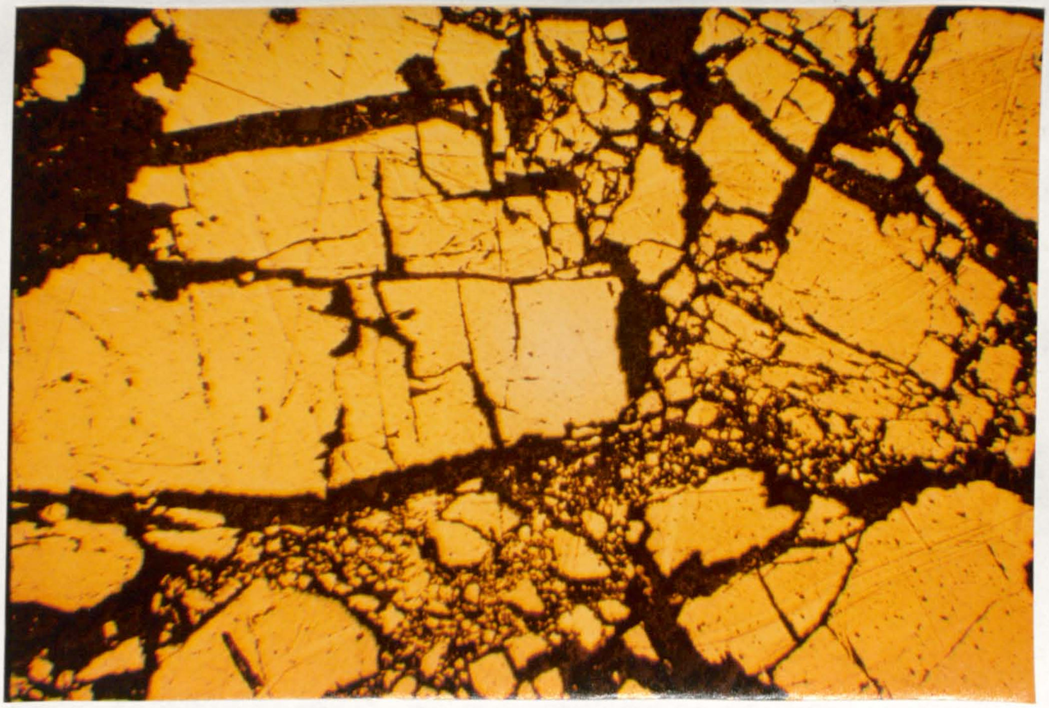


FIG. 8.1. Brittle fracture in pyrite. PPL. Air. X10.

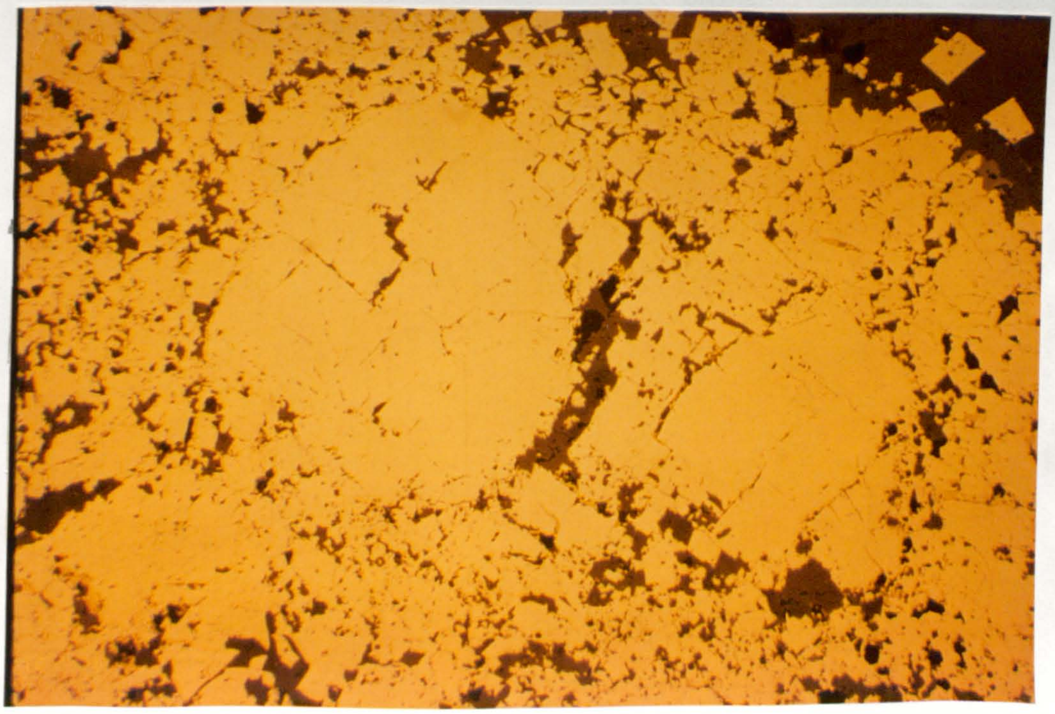


FIG. 8.2. Rolled pre-tectonic pyrites in a matrix of post-tectonic pyrite. PPL. Air. X25.

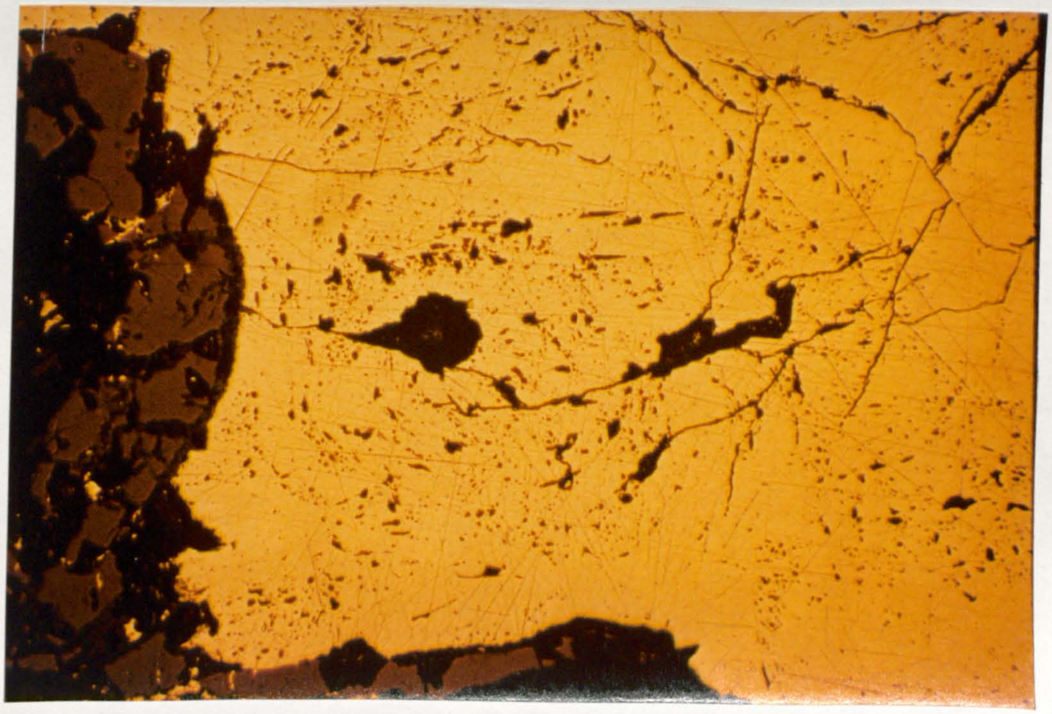


FIG. 8.3. Spiral inclusion trails in syn-tectonic pyrite with later post-tectonic overgrowth. PPL. Air. X64.

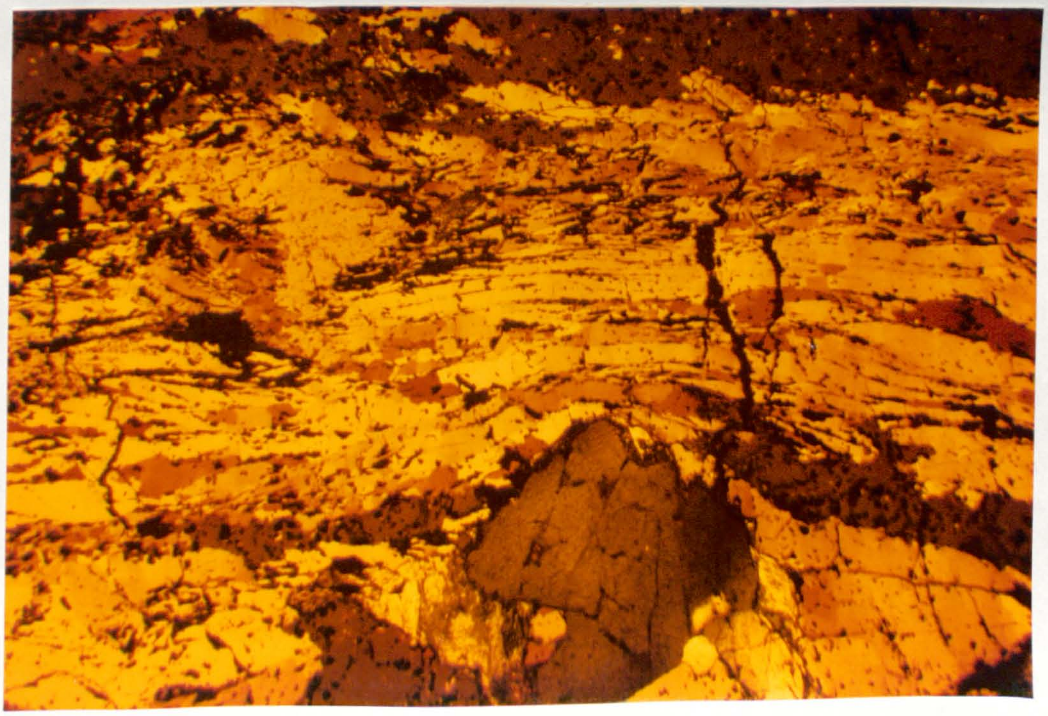


FIG. 8.4. Gneissic texture developed in pyrrhotite, bent around pre-tectonic rolled sphalerite. XPL. Air. X10.

relict gneissosity. This is a feature of the relative crystalloblastic force of the two sulphides rather than the later growth of pyrite. Pyrrhotite also forms large xenoblasts which occur within the pressure shadows of resistant crystals such as pyrite (Figs. 8.4, 8.6). A certain amount of plastic flow has also occurred and pyrrhotite occasionally infills cracks in pyrite. Sphalerite adopts two main habits dependent on the degree of stress it has undergone. Highly sheared specimens within the gneissic bands have been drawn out into long flasers (Figs. 8.4, 8.6). Other crystals have behaved much like pyrite and formed resistant bodies. Chalcopyrite has behaved very plastically and flowed to infill many cracks in the more brittle sulphides. Silicate rock xenoliths are common within the massive ore and the two usually exhibit a typical 'durchbewegung' fabric. The silicate inclusions have been extensively rolled to produce almost spherical bodies up to 10 cm across within the sulphide matrix. The silicate inclusions are usually of amphibolitic composition although rarer xenoliths of gneiss do occur.

8.3 DEFORMATION AND RECRYSTALLIZATION HISTORY OF THE ORE DEPOSIT

Response to deformation was partly a function of the composition of the ore: pure pyritic ore reacted differently to pyrite mixed with softer sulphides or with silicates. The history is complex and involves several phases of growth and deformation, but the development of the various pyrite habits provides the best evidence for the succession of events that affected the whole deposit. In the massive pyritic ore, several generations can be distinguished. The earliest

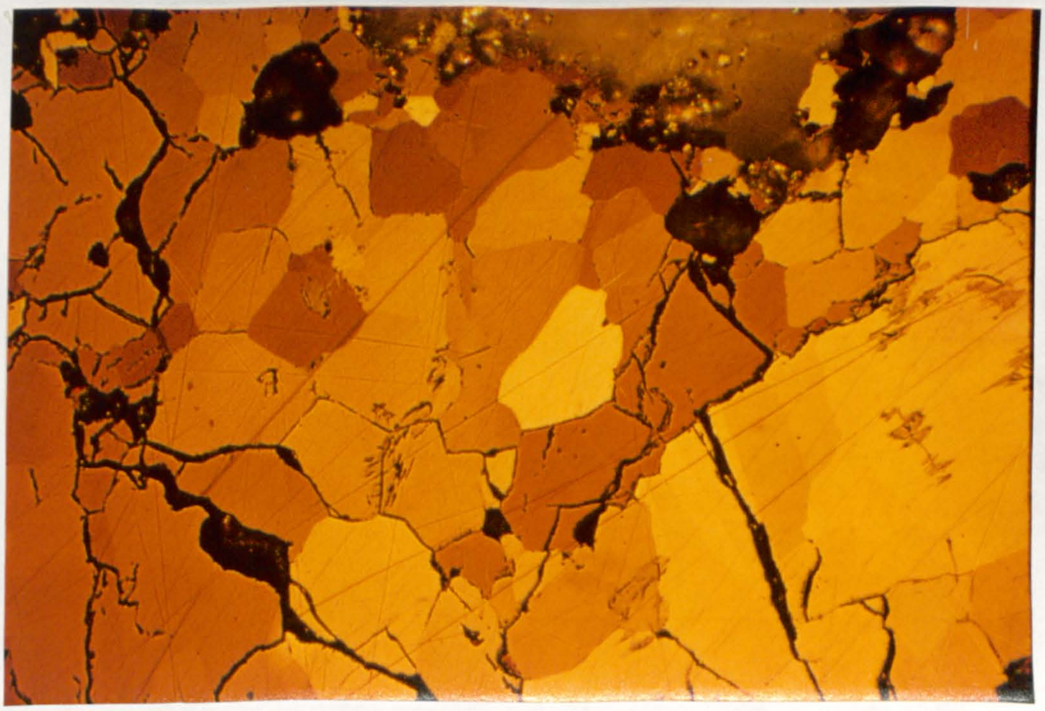


FIG. 8.5. Post-tectonic recrystallization of pyrrhotite. XPL. Oil. X64.

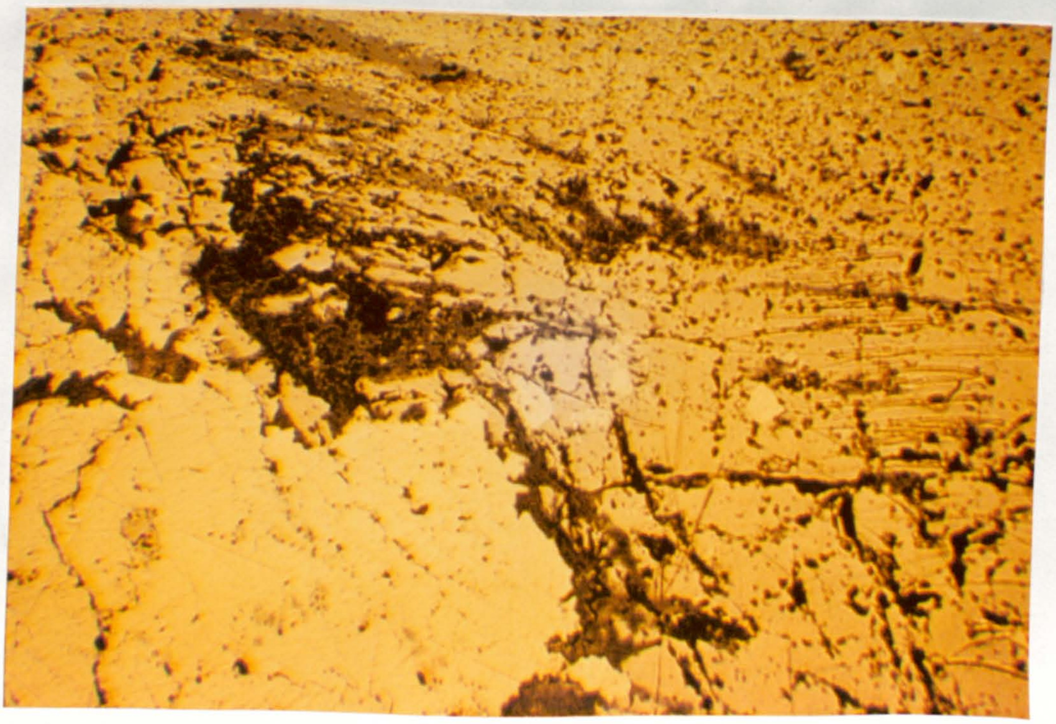


FIG. 8.6. Pyrrhotite-sphalerite 'foliation' flexed around porphyroblast of pyrite. Larger pyrrhotite grains have developed in the pressure shadow. PPL. Air. X10.

are large pre-tectonic porphyroblasts which are broken or occasionally rolled. They occur between more intensely sheared zones which contain irregular granulated fragments usually set within a xenoblastic mass of dusted and poikiloblastic pyrite (discussed later). Where softer sulphides such as pyrrhotite are present there is usually a foliation developed, which is bent around the cubes and fragments. In the most highly sheared zones, the pyrite was mylonitized. Due to the high temperature involved during the first shear episode, the pyrite in these zones annealed itself simultaneously. The result was a very irregular xenoblastic mass containing no inclusions or dusting (Figs. 8.7, 8.8). Other pyrite cubes away from the zones of shear were rolled and oval cross-sections were produced. Thus, a sequence of increasing deformation of the early pyrites can be determined: cracking and rolling progressed to granulation and finally mylonitization and recrystallization. Some syn-tectonic deformational growth also occurred with the formation of rotational crystals with spiral inclusion trails.

Overprinted on this fabric are the effects of post-tectonic pyrite growth and recrystallization of the softer sulphides. Small euhedral cubes cross-cut silicate and sulphide foliations. Where these cubes have formed in a silicate matrix they are poikiloblastic, but within a sulphide matrix they are usually non-poikiloblastic (Fig. 8.9). Occasionally elongated crystals are present in the foliation and represent a static growth which mimics the gneissic fabric. Post-tectonic xenoblastic masses (very similar in outline to the mylonitic recrystallization) commonly occur as overgrowths on existing pyrites, on their comminuted relics or interstitially between cubes. The masses are usually very dusted and poikiloblastic, with inclusions of the

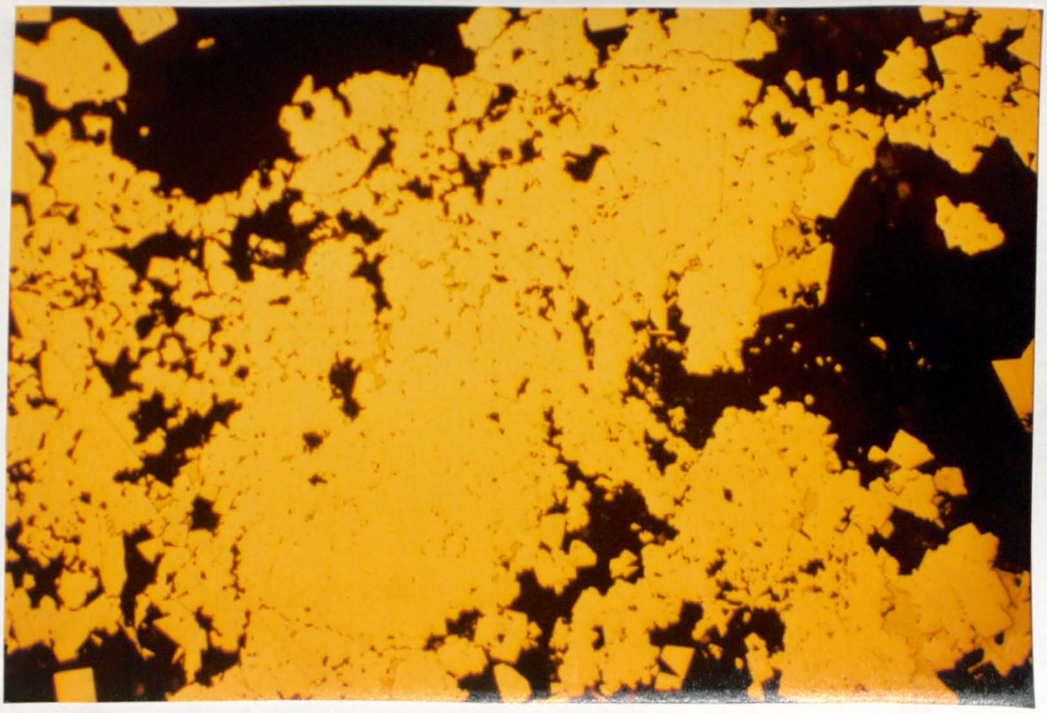


FIG. 8.7. Annealed mylonitic pyrite. PPL. Oil. X64.

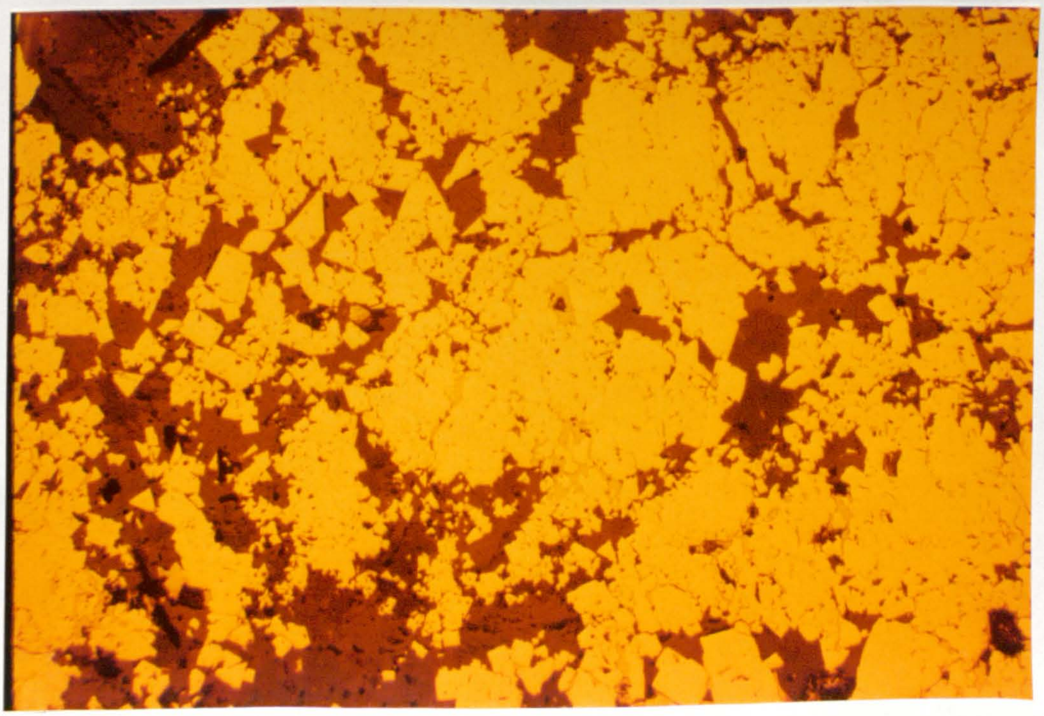


FIG. 8.8. Massive, annealed mylonitic pyrite with post-tectonic pyrite cube development. PPL. Air. X25.

FIG. 8.9. Post-tectonic poikiloblastic and non-poikiloblastic pyrites. PPL. Air. X25.

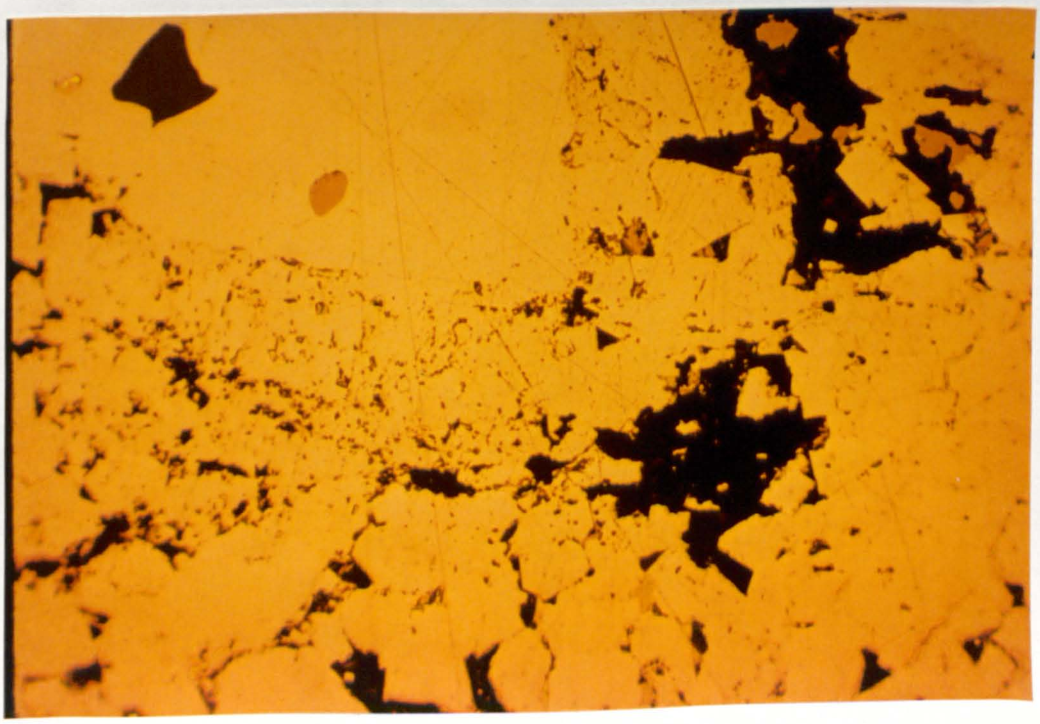


FIG. 8.10. Secondary 'dustbin' pyrite overgrowth on pre-tectonic rolled pyrite (top left). PPL. Oil. X160.

groundmass over which they have grown. The result is that a 'dustbin' pyrite has developed (Fig. 8.10). The mylonitized-annealed pyrites generally have no inclusions and do not enclose euhedral cubes. However, static post-tectonic pyrite cubes often develop on the periphery of the masses (Figs. 8.7, 8.8). Where recrystallization of the mylonite patches has been extensive, an interlocking mosaic of large crystals is formed.

A second type of 'dustbin' pyrite originates from the alteration of pyrrhotite. The alteration does not appear to be supergene since, where it occurs, there are no other oxidation products and the samples of ore are fresh. It appears therefore that the oxidation is an integral part of the metamorphism of the ore. All stages of alteration can be identified. First, pyrite veins the pyrrhotite and gives the appearance of having intruded it (Fig. 8.11). The veins then coalesce to form patches in the pyrrhotite. The replacement continues until only relics of pyrrhotite remain (Fig. 8.12). Around the margins of this secondary pyrite are fine intergrowths of magnetite produced by the alteration. The reaction can be represented as:

$$6FeS + O_2 \longrightarrow 3FeS_2 + FeOFe_2O_3 \quad .$$

This alteration explains the frequent occurrence of 'pyrite veins' (which were originally pyrrhotite) infilling cracked pre-tectonic pyrites. Some true supergene pyritization is also present in weathered specimens and classic 'birds eye' textures have developed. It is evident from the pyrite and pyrrhotite mineralogy that textures resulting from directed stress have been subsequently annealed. Evidence of the

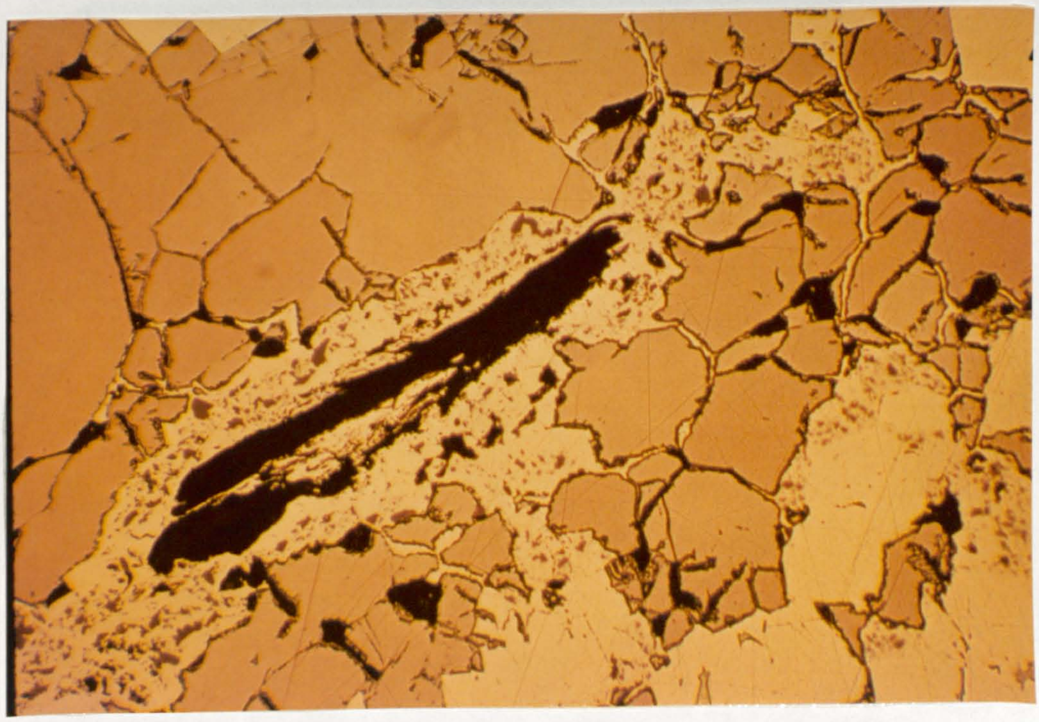


FIG. 8.11. Pyrite (yellow) and minor magnetite (grey) replacing pyrrhotite (brown). Black material is silicate. PPL. Oil. X64.

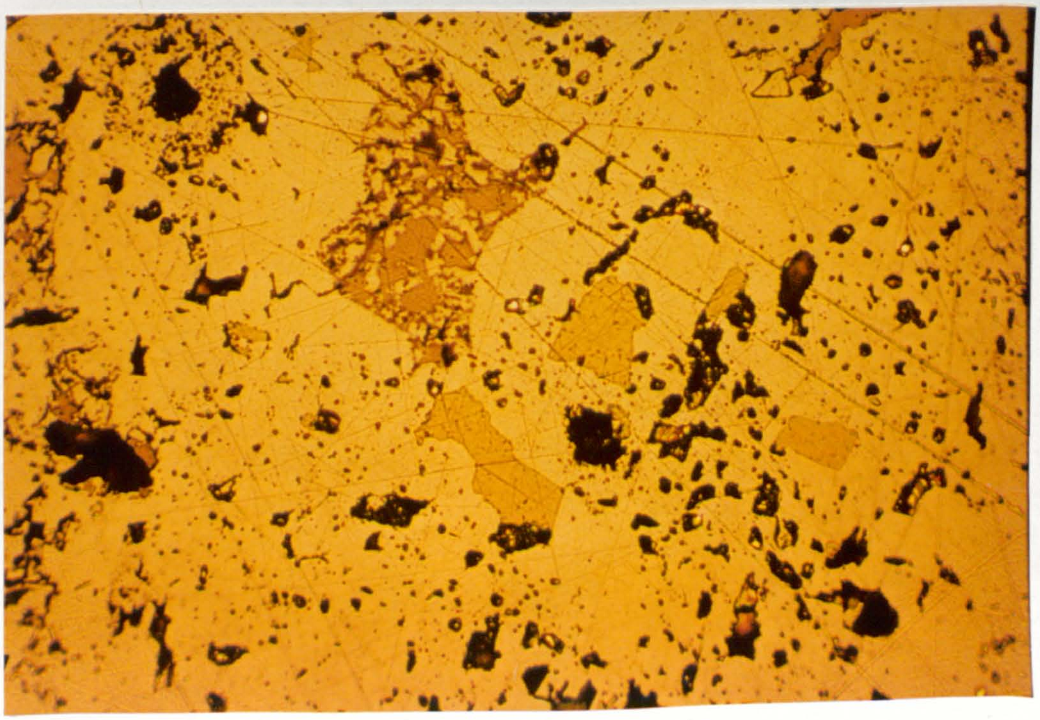


FIG. 8.12. Relict pyrrhotite (brown) and chalcopyrite (golden yellow) with pyrite (yellow) replacement. Accessory flecks of gold are also present. PPL. Oil. X160.

deformational gneissosity imparted to the ore only remains because the recrystallization of the pyrrhotite has mimicked the original lineation.

Within the mixed sulphide-silicate 'durchbewegung' rocks, shear fractures have not developed and discrete shear zones are lacking. It is suggested that these rocks were more highly plastic due to their silicate content. Thus, during the shearing there was extensive rolling of the silicate material into spheres. The pyrite present acted relatively competently, but shearing did not develop because stress was taken up by flow of the composite unit. The result was a 'durchbewegung' fabric in which the large pre-tectonic pyrites have been rolled and fractured but not sheared out.

8.4 ORE HISTORY LINKED TO SILICATE METAMORPHISM

The silicate country rocks around the Eiker mine are mid to upper amphibolite facies metamorphics with a later localised overprint of greenschist facies. The shear belt along which the mines are located shows evidence of a brittle movement at greenschist facies grade with the development of actinolite-biotite schists along shear planes. Later, high level and minor readjustments along the shears are probably linked to Oslo Graben movements. (The shear belt is only 5 km west of the major west fault of the Oslo Graben.)

There was an initial period of ore crystallization with the development of large porphyroblastic pyrites. This was followed by a period of intense shearing which resulted in rotation, shattering and granulation. In the most intensely sheared zones mylonitization and simultaneous recrystallization occurrred. This shearing was a high temperature phenomena and not only softer sulphides, but also silicate gangue flowed into cracks in the brittle minerals. Where silicate material was present (possibly removed from foot and hanging walls by the shearing) a rolled 'durchbewegung' fabric was produced. After the deformation there was a short period of recrystallization with the development of small pyrite cubes and the annealing of deformation features in the softer sulphides. All of these features in the ore were developed under amphibolite grade metamorphism. A shear zone produced in such a high grade environment would not be as sharply defined as the feature at Eiker. However, a later greenschist grade shear, resulting in brittle fracture of the silicate rocks and ore has produced a sharpening of the shear lineament.

The absence of any strong deformational features (such as kink bands or glide twins in many pyrrhotite grains) indicates a partial annealing. However, the presence of straining and lack of triple junctions also suggests that total equilibrium was not reached in recrystallization. This annealing probably occurred during the cooling period after the brittle greenschist grade shearing. The pyritization of pyrrhotite may well have occurred at this stage since there are no deformation features within the pyrite produced.

Table 8.1 shows the origins of the various types of pyrite. A correlation with the metamorphism in the silicate country rocks is shown in Table 8.2. The high grade shearing episode is considered to be contemporaneous with the formation of the mylonite zone in the west of the Kongsberg series. The deformation features of the ore body are thus of Sveconorwegian age.

TABLE 8.1

FEATURE	ORIGIN
Cracked and broken porphyroblasts	Pre-tectonic. In zones of less intense shearing, or resistant blocks in relatively incompetant silicate matrix.
Granulated cubes	Pre-tectonic. In zones of increased shearing.
Xenoblastic patches (clear of inclusions)	Pre-tectonic with syn-tectonic annealing. In zones of most intense shearing (i.e. mylonite).
Rounded pyrites	Pre-tectonic. Rotated and rounded during shearing episode.
Rounded pyrites with helicitic inclusion trails	Syn-tectonic. Inclusion trails formed during rotation through shearing.
Xenoblastic patches ('dustbin')	Post-tectonic. Two origins 1. Interstitial or secondary rim growth. 2. Pyritization of pyrrhotite.
Small cubes	Post-tectonic. Growth over silicate matrix → poikiloblastic. Growth over sulphide matrix → non-poikiloblastic.

TABLE 8.2

SULPHIDES	SILICATE ROCKS
Pre-tectonic crystallization	Sveconorwegian orogeny
High grade deformation. Production of gneissic, mylonite and 'durchbewegung' fabrics.	Mid amphibolite facies metamorphism Shear belt formed, contemporary with mylonite zone at west margin of Kongsberg Series.
Post-tectonic recrystallisation.	Decrease in metamorphic grade.
Low grade brittle deformation. Straining and deformation features (glide twins, kink bands, etc.) produced in softer sulphides.	Limited greenschist facies metamorphism. Formation of open folds and bending of supracrustals around basic bodies. Renewed movement of shear belt.
Partial annealing of above textures and pyritization.	Decrease in metamorphic grade.
	High level Oslo Graben movements. Possible movement of shear belt.

CHAPTER NINE

CHEMISTRY OF THE SILICATE ROCKS AT EIKER

9.1 INTRODUCTION

The problems of the origins of the silicate rocks are essentially the same at Eiker as at Grøsli. However, the rock types differ in their chemistry, from those of Grøsli. The supracrustal rocks are mineralogically more variable in that they may contain sillimanite, cordierite and they also contain bands of more basic material in the form of amphibolites or hornblende gneisses. The supracrustals are intruded by later basic rocks, as at Grøsli, but the intrusives are totally amphibolitized, leaving no original igneous textures.

9.2 THE CONCORDANT BASIC ROCKS OF THE SOUTHERN SUB-AREA

The relationships of these rocks with the acidic supracrustals and the basic stocks of the northern sub-area is discussed in terms of relative age and origin.

The occurrence of interbanded amphibolites and more acidic rocks is fairly common. Many workers, including Poldervaart (1953) and Wilcox and Poldervaart (1958), suggested a primary sedimentary banding as the reason for the interbanding. However, Evans and Leake (1960) studied the Connemara striped amphibolites and suggested that the basic bodies were igneous, for several reasons. Their average analyses were very similar to alkaline-olivine basalt analyses, and plots of

Niggli mg versus Niggli c, and al-alk versus c, produced igneous rather than sedimentary trends. They also stated that, to achieve the correct basic composition from sedimentary rocks, a mixture of Na-rich greywackes and Fe-rich dolomites was required; two peculiar rock types not present in the area.

Various other chemical trends such as Niggli mg versus Cr ppm also gave typical igneous trends. Evans and Leake concluded that the amphibolites were formed from either intrusive or extrusive igneous rocks. Subsequent work (e.g. Van de Kamp, 1969 and Stephenson, 1973) indicated that many amphibolites were in fact igneous in origin. Van de Kamp stated that nearly all amphibolites were derived from basic igneous rocks with a few formed from metasomatized sediments.

Orville (1969) showed that three mechanisms could produce amphibolites; (a) recrystallization of basic intrusive or extrusive igneous rock, (b) recrystallization of a marl or carbonate bearing shale, (c) recrystallization of a parent sedimentary rock with metasomatic addition and/or subtraction of significant amounts of non-volatile material. He stated that any hornblende-plagioclase (amphibolite) composition, by virtue of the hornblende crystal chemistry, will fall within the basalt fields or on the basaltic differentiation trends (such as those used by Leake). He then suggested that attention should be directed towards processes which would produce a hornblende-plagioclase assemblage. Orville considered that metamorphism of carbonate-shale mixtures rarely produced amphibolites, since none of the sedimentary trends suggested by Leake (1964) were commonly developed, but that metasomatized sediments could plot on igneous differentiation trends.

Taking a hypothetical carbonate-rich shale enclosed in a host of carbonate-poor shale, he suggested that metamorphism would cause reaction, with the leaching of excess $CaCO_3$ from the carbonate rock and reciprocal exchange of Ca, Mg and Fe cations between the two layers. As a result of this reaction, certain features could be expected to be present in the amphibolites and their surrounding rocks. There would be a close and systematic relationship between interlayered amphibolites, marbles, pelitic schists, calc-silicate layers and gneisses. Calc-silicate or carbonate-bearing layers might be present within the amphibolite. The calc-silicate layers represent intermediate compsotions between the two end member assemblages. Marbles might not be present if the original carbonate layers contained less than 40 percent $CaCO_3$.

The country rocks adjacent to the amphibolite would be expected to have a higher Ca content, but lower Mg and Fe content than those distant from the amphibolites: muscovite should be totally absent since this would be the first mineral to disappear in the reactions. These final statements suggest that this process has not caused formation of the amphibolites at Eiker. Attendant acidic gneisses are often rich in muscovite, while their Ca content is no higher than samples remote from the amphibolite bands.

Smithson et al (1971) concluded that only small amounts of amphibolite could be produced with any certainty by the method suggested by Orville. Thus, since the rigid requirements of Orville's model are not met at Eiker, it appears that the concordant amphibolites are not metasomatic in origin.

Figs. 9.1 and 9.2 are plots of Niggli mg versus c and 100mg/c/al-alk respectively. As with the plots of basic rock analyses from Grøsli,

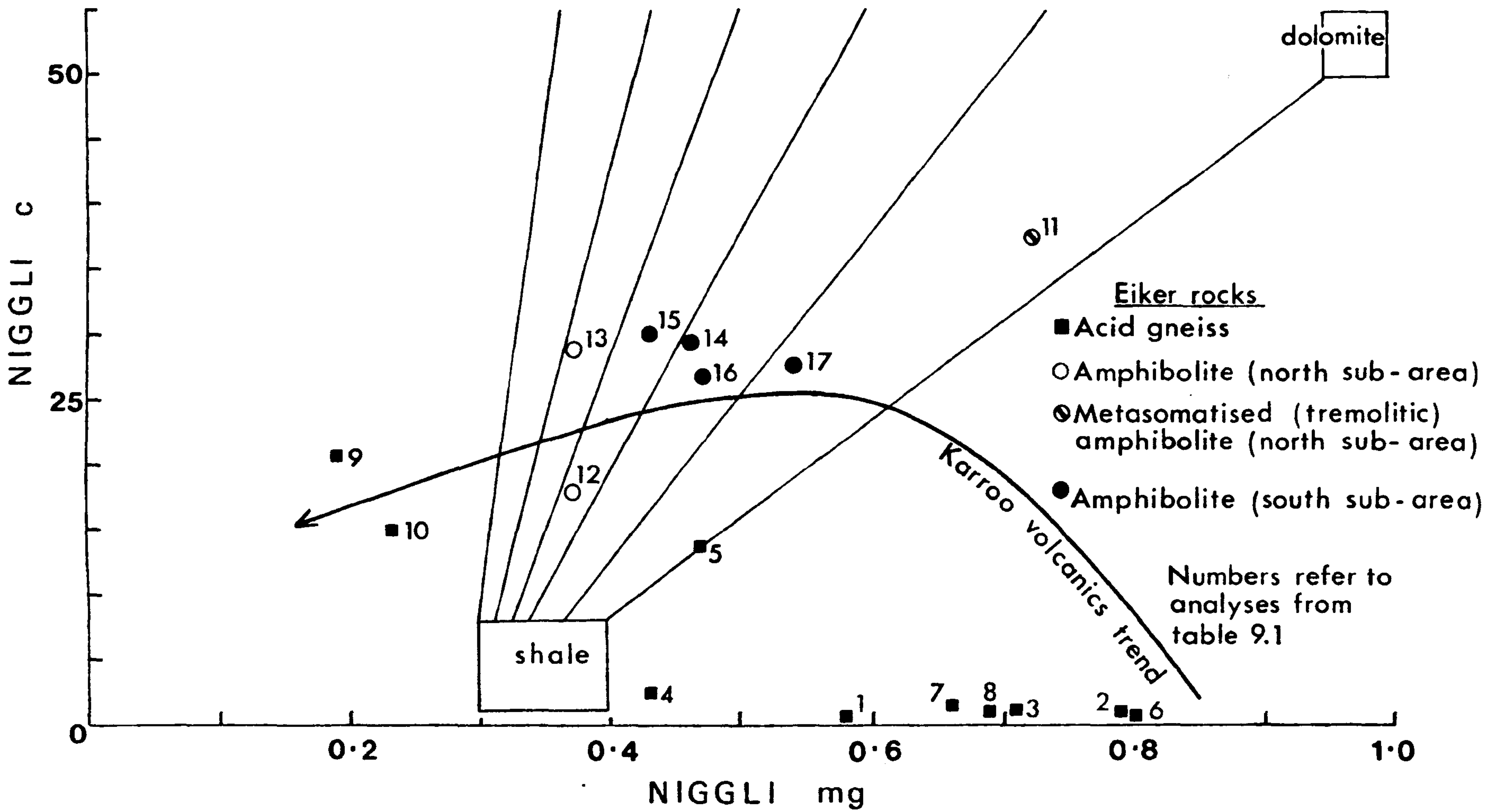


FIG. 9·1 (after Leake, 1964)

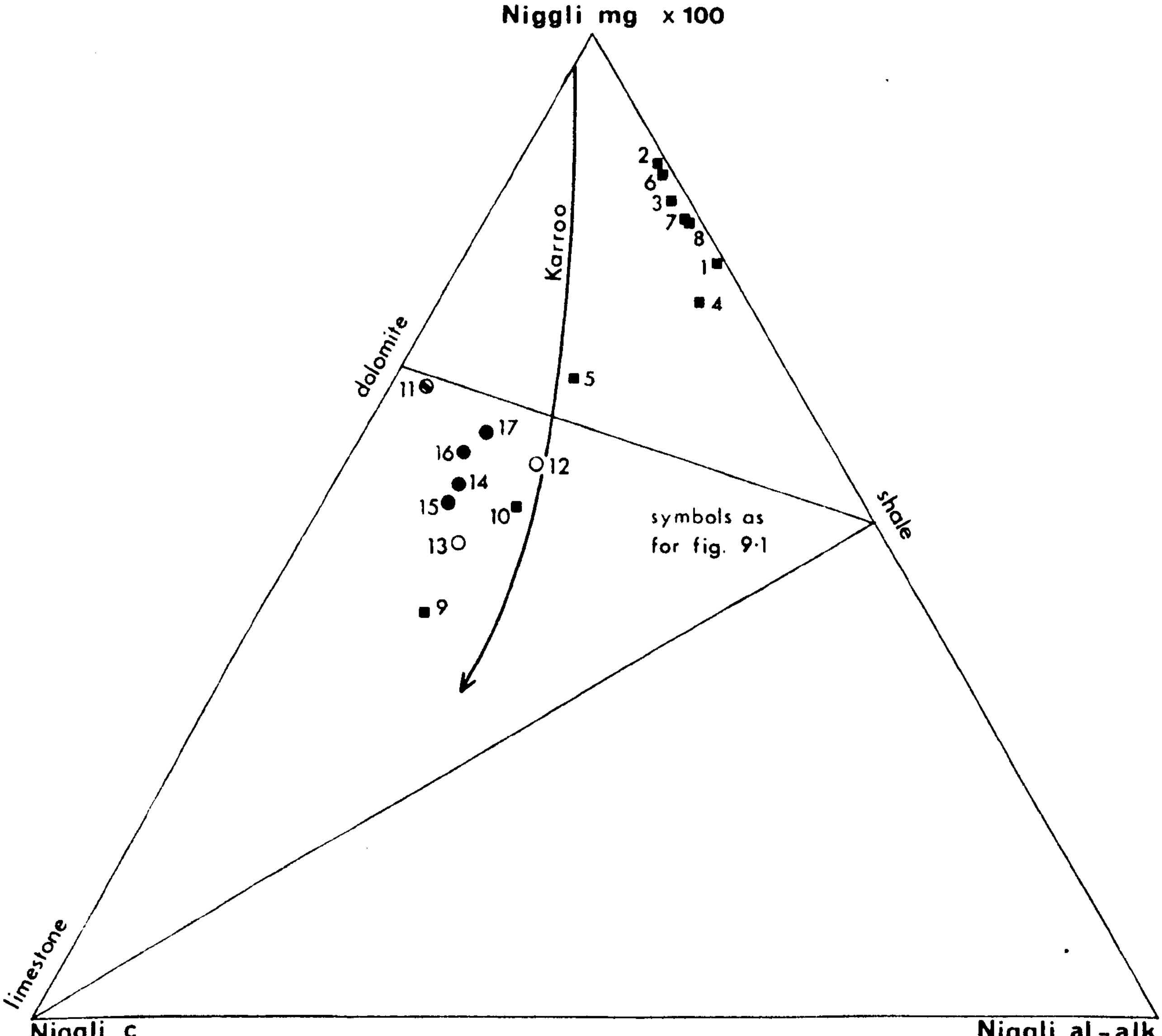


FIG. 9·2 (after Leake, 1964)

Figs. 9.1 and 9.2 show that the dispersal of points (nos. 12 to 17) and overlap of igneous and sedimentary fields render the diagrams of limited use. Analyses of intermediate and acidic supracrustals have also been plotted (nos. 1 to 10). These two diagrams indicate that two groups of supracrustal gneisses are present; a large group with very low Niggli c values (nos. 1 to 4 and 6 to 8), and a pair of analyses (nos. 9 and 10) which have higher Niggli c values. Analysis 5 appears to be isolated in an intermediate position between these groups in the two diagrams. A plot of Niggli si versus Niggli mg (Fig. 9.3) shows the same division of analyses as the previous diagrams and indicates that the basic rocks and the larger group of gneisses have negative (igneous) trends.

9.3 THE INTERMEDIATE AND ACIDIC SUPRACRUSTALS

The most striking feature of all but two of the gneisses in the Eiker region is their high MgO value (Table 9.1). These values put the rocks well out of the range of the common igneous rocks arkoses and greywakes. They are also higher in MgO content than most non-dolomitic shales and related rock types (Table 9.3).

Two of the analyses (nos. 9 and 10) show much lower MgO values and have bulk chemistries similar to those in the rhyo-dacitic field of Le Maitre (1976) (Table 4.4). These differences explain the two groups of gneisses on Figs. 9.1 and 9.2. Analysis 5 (the most silica-poor gneiss) plots well within the amphibolite grouping on Fig. 9.3. It also plots on a tie-line between shale and dolomite analyses on Fig. 9.1. This rock appears to be chemically distinct from the more

TABLE 9.1. MAJOR ELEMENT CHEMISTRY OF ROCKS IN THE EIKER AREA (wt. %)

ANALYSIS	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	TOTAL (excluding H_2O)
1	70.66	0.38	13.55	2.26	5.16	0.13	5.63	0.13	0.12	1.61	99.63
2	65.10	0.29	12.80	4.11	2.34	0.09	12.71	0.29	0.27	1.33	99.33
3	72.50	0.26	11.50	0.84	4.32	0.09	6.87	0.25	0.23	2.21	99.07
4	76.00	0.20	9.14	1.19	7.70	0.13	3.79	0.41	0.07	1.16	99.79
5	54.28	0.50	17.80	1.99	11.47	0.47	6.72	4.22	1.98	1.53	99.96
6	66.40	0.34	12.81	0.12	5.09	0.09	12.05	0.21	0.19	1.15	98.45
7	75.45	0.26	11.27	1.17	4.40	0.08	5.94	0.15	0.18	2.12	100.02
8	70.00	0.31	13.03	0.42	5.18	0.06	7.18	0.25	0.23	1.75	98.41
9	67.81	0.78	12.87	0.95	4.77	0.11	0.76	3.05	3.53	3.20	97.83
10	68.34	0.35	14.24	2.66	3.34	0.12	1.10	2.23	5.25	1.01	98.64
11	49.05	0.30	7.70	1.30	7.79	0.14	13.13	16.72	0.71	0.63	97.47
12	56.87	0.76	15.30	1.74	11.38	0.45	4.37	4.85	2.86	0.01	98.59
13	53.26	0.67	16.35	2.46	9.75	0.19	4.05	8.52	2.26	0.82	98.33
14	57.19	0.95	15.60	1.85	6.09	0.15	3.80	7.42	3.45	1.49	97.99
15	57.69	0.85	16.76	1.79	5.83	0.13	2.96	7.11	4.12	1.76	98.50
16	56.17	0.98	14.99	1.64	7.36	0.15	4.41	7.14	3.20	2.10	98.14
17	48.57	1.46	16.74	1.81	9.97	0.19	7.83	9.79	2.25	0.11	98.72

TABLE 9.1 continued

Anals. 1 to 8.	Mg-rich gneisses	- Southern sub-area
Anals. 9 and 10.	Gneisses	- Southern sub-area
Anals. 11 to 13.	Amphibolites	- Northern sub-area
Anal. 11.	Metasomatized amphibolite	
Anals. 14 to 17.	Amphibolites	- Southern sub-area

TABLE 9.2. TRACE ELEMENT CHEMISTRY OF ROCKS IN THE EIKER AREA (ppm)

ANALYSIS	Cr	V	Co	Ni	Cu	Zn	Cd	Rb	Sr	Y	Zr	Ba	Ce
1	<50	10	7	7	4	98	N.D.	19	5	19	60	355	30
2	<50	12	6	9	5	242	N.D.	21	21	9	74	165	4
3	<50	8	6	7	5	127	N.D.	24	8	10	55	379	14
4	<50	12	14	10	12	64	N.D.	11	4	22	168	278	12
5	131	264	32	30	2026	880	50	2	40	17	35	1681	29
6	50	11	6	8	5	204	N.D.	14	17	12	73	98	8
7	50	9	8	7	41	127	N.D.	32	10	6	51	501	N.D.
8	50	11	10	9	3	110	N.D.	14	13	15	74	318	24
9	50	30	13	6	16	92	N.D.	49	229	53	398	1466	52
10	50	40	9	8	15	61	N.D.	5	254	12	44	216	27
11	1730	127	62	341	N.D.	45	83	8	152	7	26	46	63
12	101	299	52	18	N.D.	233	29	1	109	19	30	72	33
13	97	295	38	21	141	99	59	4	362	20	19	316	34
14	134	200	27	31	88	121	3	35	355	24	66	370	15
15	85	138	26	23	65	88	3	27	414	36	114	617	34
16	157	237	32	34	145	121	N.D.	49	349	30	146	569	27
17	201	186	67	125	152	120	69	3	222	33	102	180	23

N.D. = Not detected.

TABLE 9.3

ANALYSES OF SHALES AND RELATED ROCKS

	1	2	3	4
SiO_2	66.87	69.96	73.71	52.00
TiO_2	0.47	0.59	0.50	-
Al_2O_3	15.36	10.52	7.25	16.11
Fe_2O_3	2.81	3.47	2.63	4.69
FeO	1.89		0.44	
MnO	0.05	0.06	-	-
MgO	2.40	1.41	1.47	4.10
CaO	0.34	2.17	1.72	8.26
Na_2O	1.21	1.51	1.19	2.76
K_2O	6.60	2.30	1.00	1.74
H_2O	1.35	5.74	9.82	9.64
P_2O5	0.23	0.18	0.24	-
CO_2	0.28	1.40	-	-
SO_3	-	0.03	0.16	0.09
Cl	-	0.30	-	-
C	0.04	0.66	-	-

1. Precambrian argillite, Michigan, U.S.A.
2. Av. of 235 samples (muds and silts), Mississippi delta, U.S.A.
3. Diatomaceous shale, California, U.S.A.
4. Varved clay, Canada.

All from Pettijohn (1957)

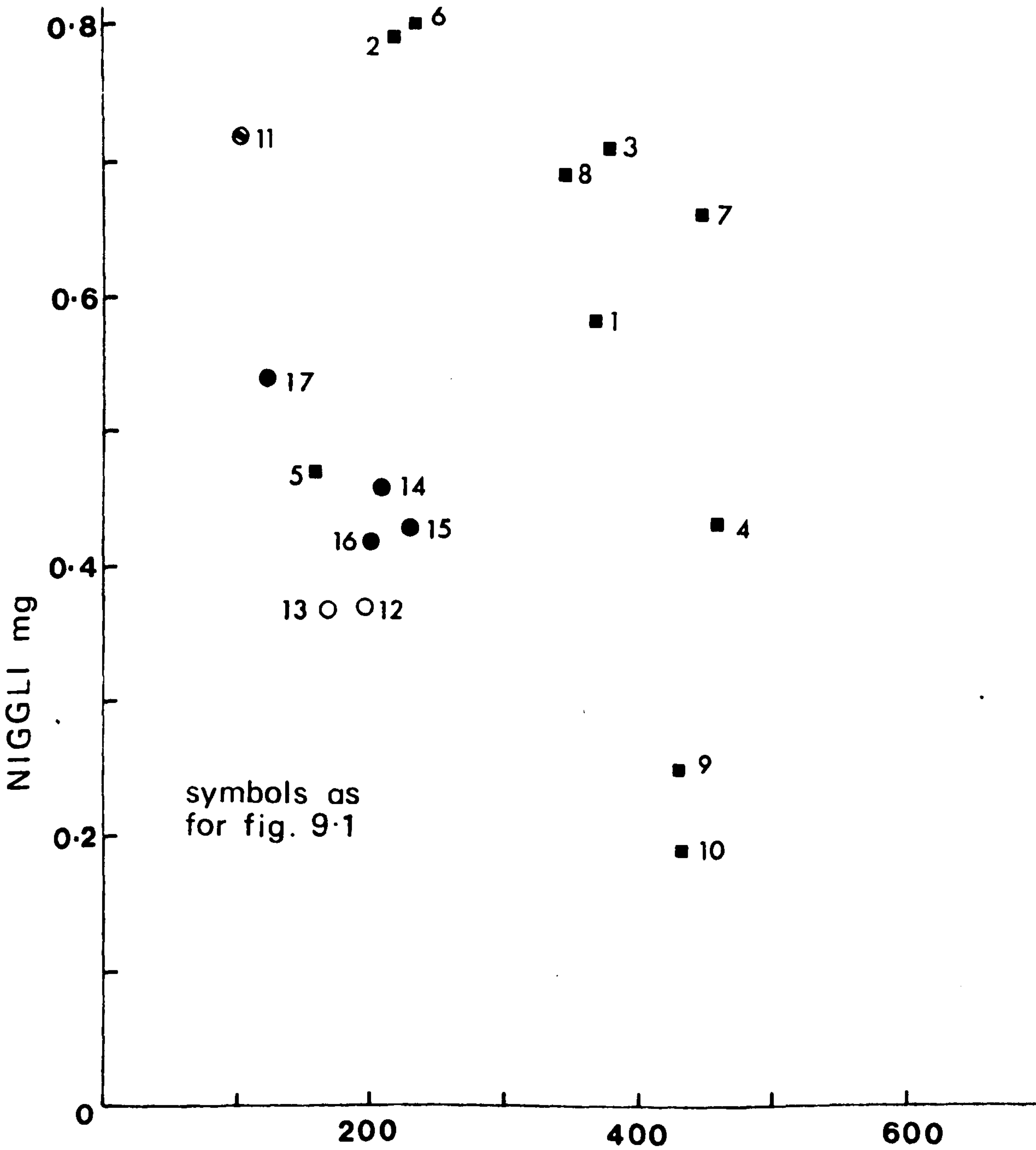

0·8
0·6
0·4
0·2
0
200
400
600
NIGGLI si
NIGGLI mg
symbols as
for fig. 9·1
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17

FIG. 9·3

acidic gneisses, while its isolation from the amphibolite groupings in Figs. 9.1 and 9.2 suggest a different affinity than that of the basic rocks. It could be considered a pelite, rich in ferromagnesian components which, on metamorphism, produced a garnet-hornblende-(quartz-muscovite) assemblage.

The negative slope of Fig. 9.3 for the MgO rich gneisses (nos. 1 to 8) is the opposite to that which Van de Kamp et al (1976) considered to be diagnostic of sediments. Accordingly, these Mg-rich rocks could have had an igneous origin. They may be volcaniclastic deposits or hydrothermally altered lavas. This idea is further considered in Chapter Eleven in terms of the total environment of ore deposition: it is concluded that the rocks are, in fact, most likely to represent the metamorphosed equivalents of volcaniclastics or altered lavas. This conclusion is based on the major element differences between the Mg-rich rocks (1 to 8) and the two analyses (9 and 10) which most resemble acidic lavas. The gneisses at Eiker, therefore, could be a suite of altered acidic lavas interbedded with basic igneous units (tuffs, lavas or intrusives). Sillimanite bearing rocks may have the same origins or may represent interbedded sediments.

9.4 THE BASIC INTRUSIVES OF THE NORTHERN SUB-AREA

The amphibolitized stocks in the northern sub-area are obviously intrusive. Two of the analyses (nos. 12 and 13) plot with a wide spread either side of the Karroo differentiation trend and to the more highly differentiated end than the southern sub-area amphibolites.

Analysis 11 plots well away from the igneous trend and appears to

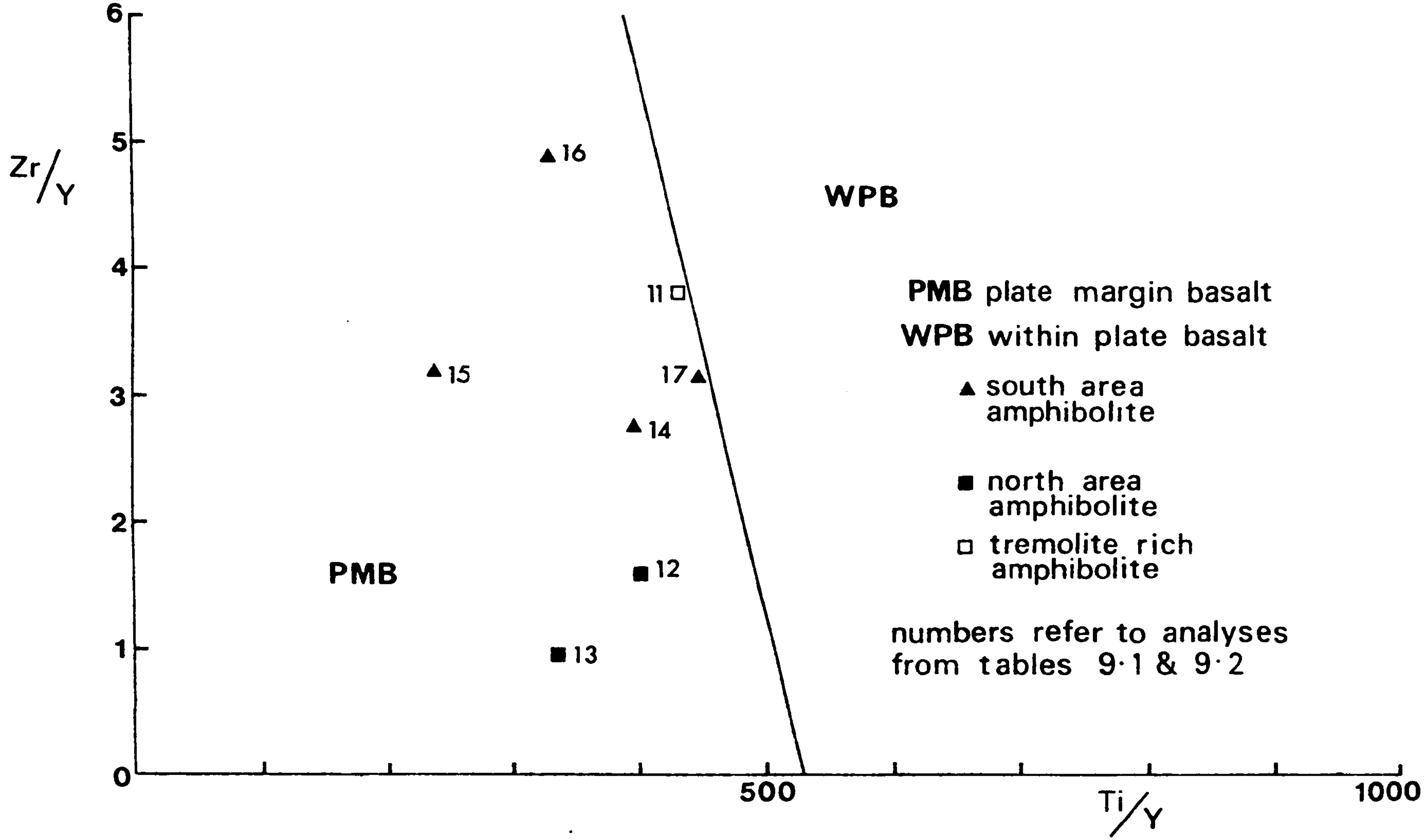


FIG. 9·4 (after Pearce and Gale, 1977)

be of Ca-rich pelitic origin. This effect has been produced by the presence of substantial amounts of tremolite in the rock which has replaced the normal hornblende metasomatically.

The effect of metasomatism can be seen in Figs. 9.1 and 9.2 reflecting the increase in Ca and Mg (and corresponding drop in Fe content) in the replacement of the hornblende by tremolite.

Figs. 9.1 to 9.3 show the difficulty of assigning origins to basic rocks on the basis of major element analysis (especially if field relationships may be open to more than one possible interpretation).

9.5 COMPARISON OF THE BASIC ROCKS IN THE NORTHERN AND SOUTHERN SUB-AREAS

Several plots have been used to show the differences in chemistry between the basic rocks in the northern sub-area and those in the southern sub-area. Fig. 9.4 indicates that both groups are plate-margin basalts (PMB). Fig. 9.5 separates three of the rocks from the southern sub-area (nos. 14, 16 and 17) as mid-ocean ridge basalts (MORB), but the remaining analysis (no. 15) and two from the northern sub-area (nos. 12 and 13) are around the island-arc basalt (IAB)-MORB boundary. The third analysis from the northern sub-area, which plots well within the MORB field is the tremolite-rich amphibolite which appears to have had its trace element ratios altered by metasomatism. This conclusion is evinced by the extremely high Cr and low Y values (Table 9.3) causing the analysis to plot away from the usual fields.

Fig. 9.6 suggests that the southern sub-area amphibolites (nos. 14 to 17) are of MORB type while those of the northern sub-area (nos. 12 and 13) are island-arc tholeiites (IAT). This is somewhat supported by

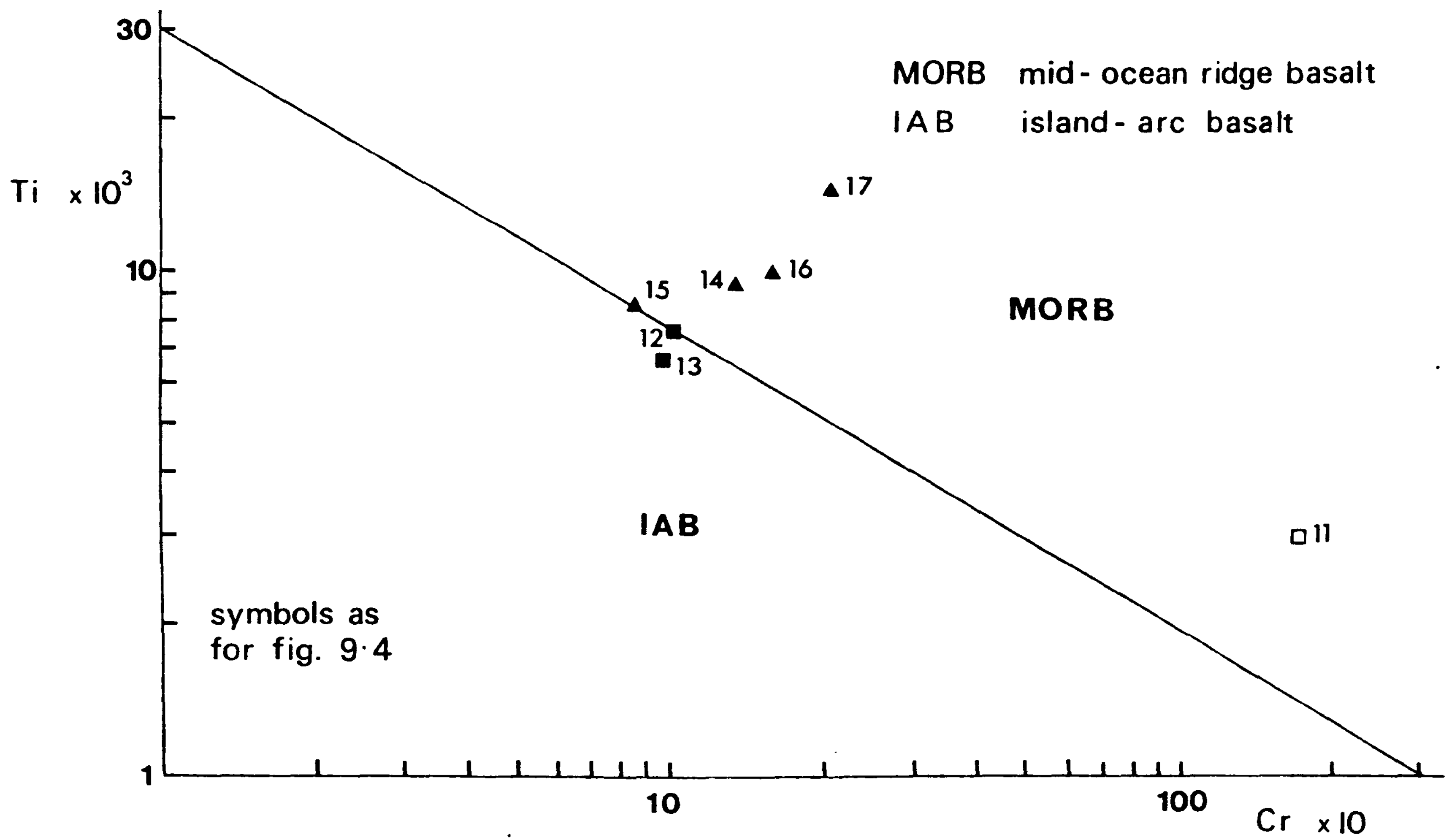


FIG.9·5 (as fig.9·4)

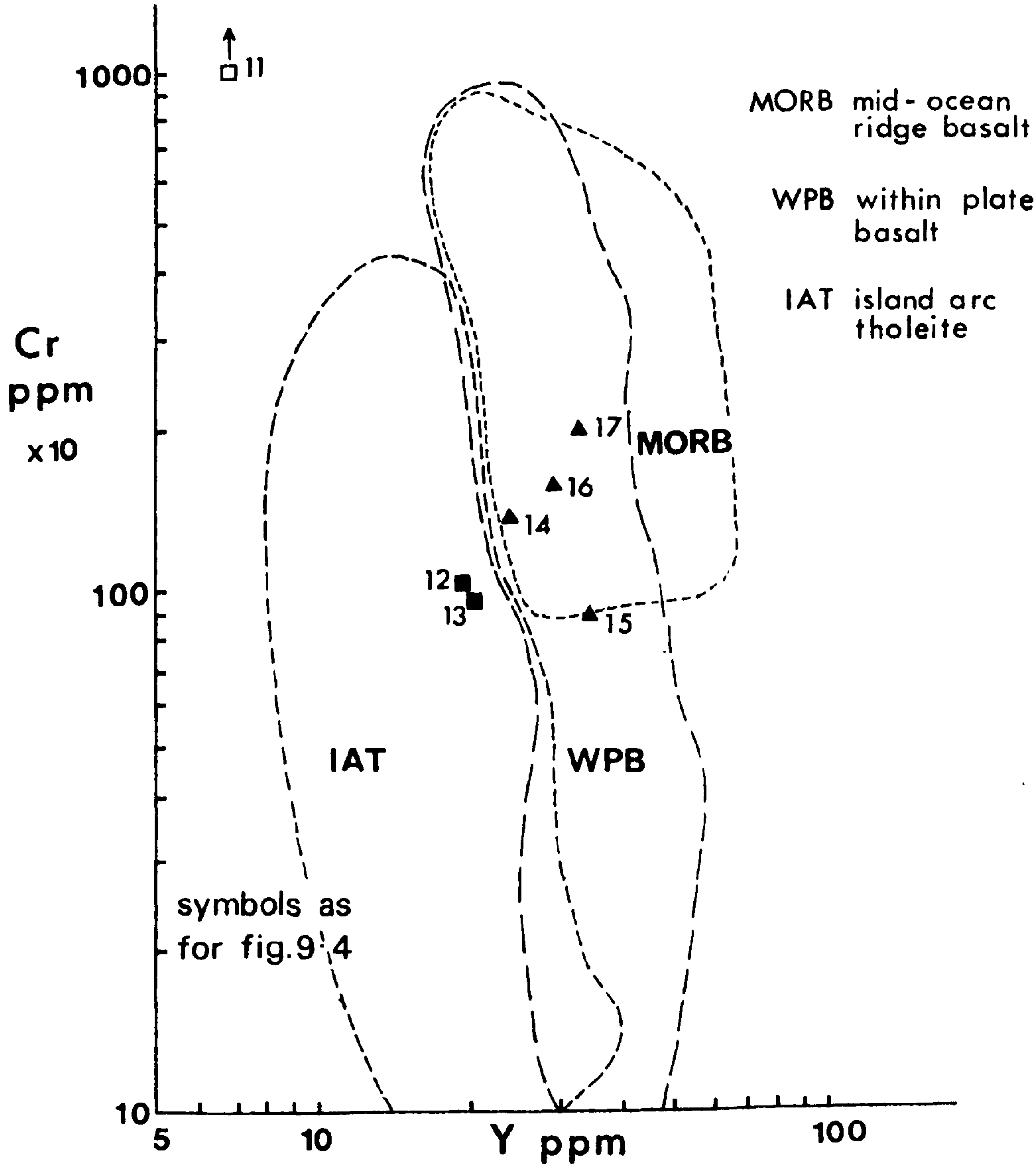


FIG.9·6 (after Pearce, in press)

Fig. 9.7 which shows a definite separation between the two groups, with the southern sub-area rocks (nos. 14 to 17) displaying higher Zr and Ti values than the northern sub-area rocks.

Thus, although the small number of analyses make conclusions difficult and speculative, there does appear to be a definite difference between the basic rocks from the two sub-areas. Those from the north appear to have island-arc characteristics while those to the south appear to have MORB features. However, in all diagrams, there is a grouping of rocks from both areas around boundaries, which may be significant and could indicate transitional magmas.

The basic rocks from both areas could be interpreted in three ways. Firstly, they may represent two different magmas, as suggested by Fig. 9.6. Secondly, they may represent different stages in the evolution of the same magma. Thirdly, they may represent one transitional magma intruded synchronously in both north and south sub-area.

Transitional MORB/IAT magma types have been described by Reay et al (1974), transitional MORB/WPB types by Pearce and Gale (1977) and transitional WPB/IAT types by De Long et al (1976). If the amphibolites are considered as the product of one magma, then this must have been of transitional MORB/IAT type. Reay et al noted these transitional basic rocks in the Tongan arc, while Pearce and Gale and Pearce (in press) describe rocks with these features from Cyprus and Oman. Pearce considers that, in the latter two localities, the magma is produced from spreading in a marginal or back-arc basin above a shallow dipping subduction zone. However, the division between the obviously intrusive northern amphibolites and the interbanded southern sub-area amphibolites coincides with the separation between IAT/IAB and MORB, respectively.

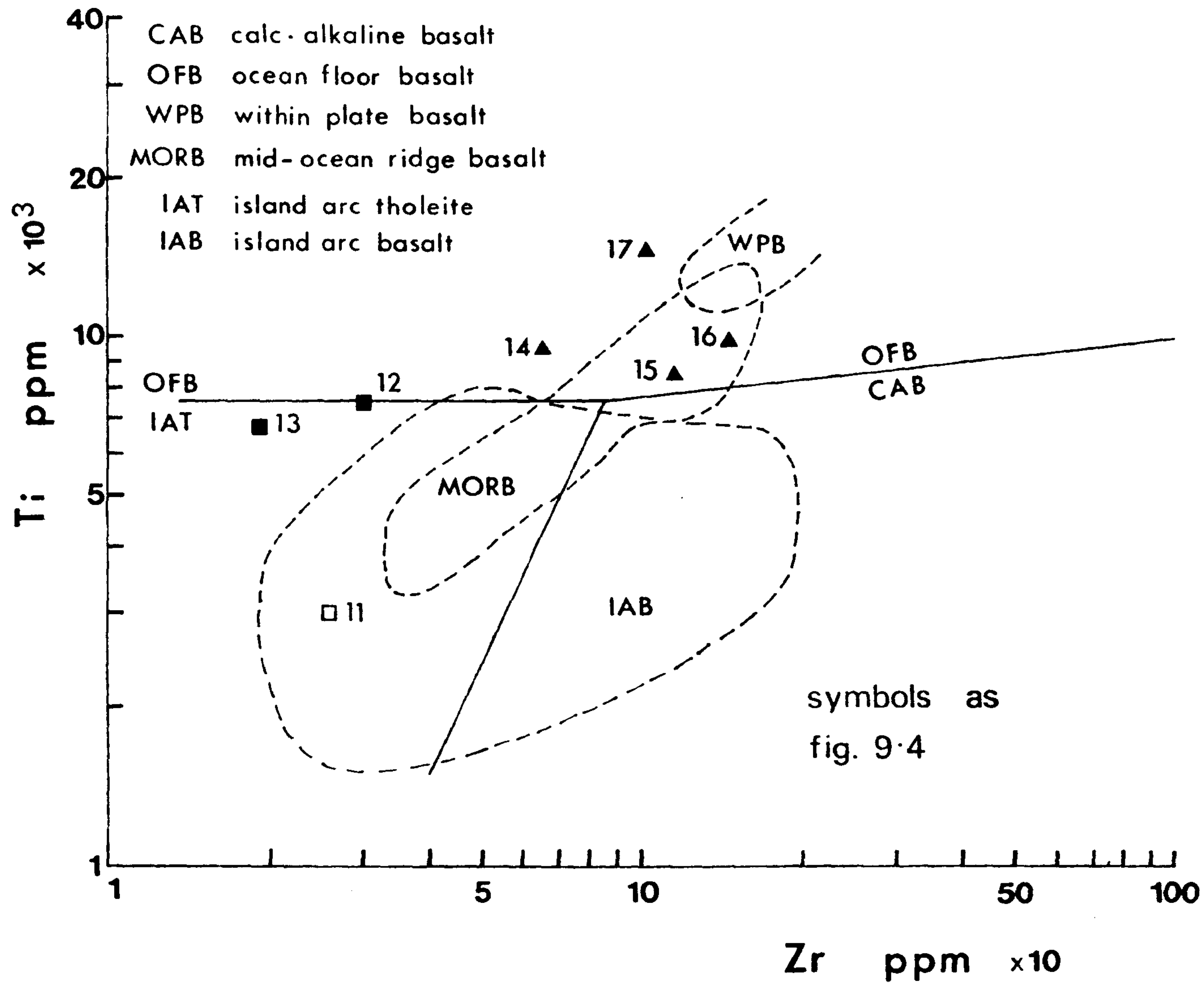


FIG. 9·7 (after Pearce and Cann, 1973)

This suggests that there may be a real difference (and possibly an age difference) between the two groups.

In the southern sub-area, there appears to be a contradiction in that basic rocks with MORB trace element values (characteristic of ensimatic rifting) occur interbanded with acidic continental crustal material. Pearce and Gale (1977) quote such an occurrence in the Røros and Finnmark areas of Norway and state that the basic rocks although showing MORB characteristics are "somewhat transitional towards within-plate basalts" (WPB) and are considered to form at continental margins. This explanation could certainly be applied to the amphibolites of the southern sub-area at Eiker. The amphibolites of the northern sub-area are again somewhat transitional in nature and lie close to the boundary in Figs. 9.5 and 9.7. These rocks could therefore be considered of transitional MORB/IAT nature.

Field relationships could favour two generations of basic rock igneous activity; an earlier phase of MORB/WPB type rock in the south and a later phase of MORB/IAT type in the north. Considering the small number of analyses, the chemical data must be viewed with caution. This is particularly the case with trace element analyses, suggesting results somewhat at variance with field relationships and major element data which are consistent with the southern sub-area amphibolites being part of the supracrustal sequence (see section 11.3.4).

CHAPTER TEN

GEOTHERMOMETRY AND GEOBAROMETRY OF THE ORE DEPOSITS AT GRØSLI AND EIKER

10.1 INTRODUCTION

Electron microprobe data giving the precise composition of pyrrhotite and the degree of iron substitution in sphalerite is presented in Tables 10.1 and 10.2. All values were taken from samples which were part of the three phase assemblage; pyrrhotite-pyrite-sphalerite and each mineral grain was in contact with grains of the other two minerals.

10.2 THE PYRITE-PYRRHOTITE GEOTHERMOMETER

Arnold (1962) experimentally established the temperature curve for crystallization of pyrrhotite in coexistence with pyrite. This univariant solvus curve related temperature to the metal-sulphur ratio of the pyrrhotite. The curve was established between 743°C (the pyrite peritectic, where pyrite decomposes to pyrrhotite and a sulphur rich liquid) and 325°C (below which structural inversions of pyrrhotite complicate experiments). Arnold estimated that "at the sulphur vapour pressure of the system", the iron content of pyrrhotite fell from 46.29 atomic percent ($\approx Fe_{10}S_{11}$) at 325°C to 45.97 atomic percent ($\approx Fe_5S_6$) at 600°C (Fig. 10.1). Confining pressures of up to 2kb did not move the solvus. Arnold stated that providing the pyrrhotite and pyrite coexisted in equilibrium and that the equilibrium pyrrhotite composition had not changed during subsequent geological events, then the method could be used to estimate temperatures of formation of naturally occurring pyrite-

TABLE 10.1

ELECTRON MICROPROBE ANALYSES OF PYRRHOTITES (ATOMIC %)

	1	2	3	4	5
Fe	47.03	47.24	46.96	46.93	46.93
S	52.97	52.76	53.04	53.07	53.07

	6	7	8	9	10
Fe	46.91	46.92	46.99	46.75	47.38
S	53.09	53.08	53.01	53.25	52.62

	11	12	13	14	15
Fe	47.30	47.38	47.43	47.42	47.18
S	52.70	52.62	52.57	52.58	52.82

	16	17	18	19	20
Fe	46.70	46.92	46.93	47.07	47.05
S	53.30	53.08	53.07	52.93	52.95

	21	22	23	24
Fe	45.59	45.56	45.53	45.58
S	54.41	54.44	54.47	54.42

1 to 4	Grøsli.	Borehole 7.	Epidote amphibolite.
5 to 9	Grøsli.	Borehole 11.	Gneiss with hydrothermal alteration.
10 to 14	Grøsli.	Borehole 4.	Gneiss with much hydrothermal alteration (gangue).
15 to 17	Grøsli.	Borehole 10.	Metagabbro.
18 to 20	Grøsli.	Borehole 6.	Gneiss with no hydrothermal alteration.
21 to 24	Eiker.		Massive sulphide ore.

TABLE 10.2

ELECTRON MICROPROBE ANALYSES OF IRON IN SPHALERITES (MOLE % FeS)

1	12.7	15	13.3
2	12.9	16	15.3
3	14.0	17	14.6
4	14.5	18	13.1
5	12.8	19	13.0
6	14.8	20	14.1
7	12.9	21	13.3
8	13.1	22	13.8
9	12.3	23	15.0
10	13.3	24	11.5
11	12.3	25	12.2
12	12.7	26	11.6
13	11.8	27	11.9
14	13.1	28	12.7

1 to 5	Grøsli.	Borehole 7.	Epidote amphibolite.
6 to 9	Grøsli.	Borehole 11.	Gneiss with hydrothermal alteration.
10 to 14	Grøsli.	Borehole 4.	Gneiss with much hydrothermal alteration.
15 to 19	Grøsli.	Borehole 10.	Metagabbro.
20 to 23	Grøsli.	Borehole 6.	Gneiss with no hydrothermal alteration.
24 to 28	Eiker.		Massive sulphide ore.

Analyses 3, 4, 6, 10, 14, 16, 17, 23, 28 are of iron rich patches within the sphalerite crystals.

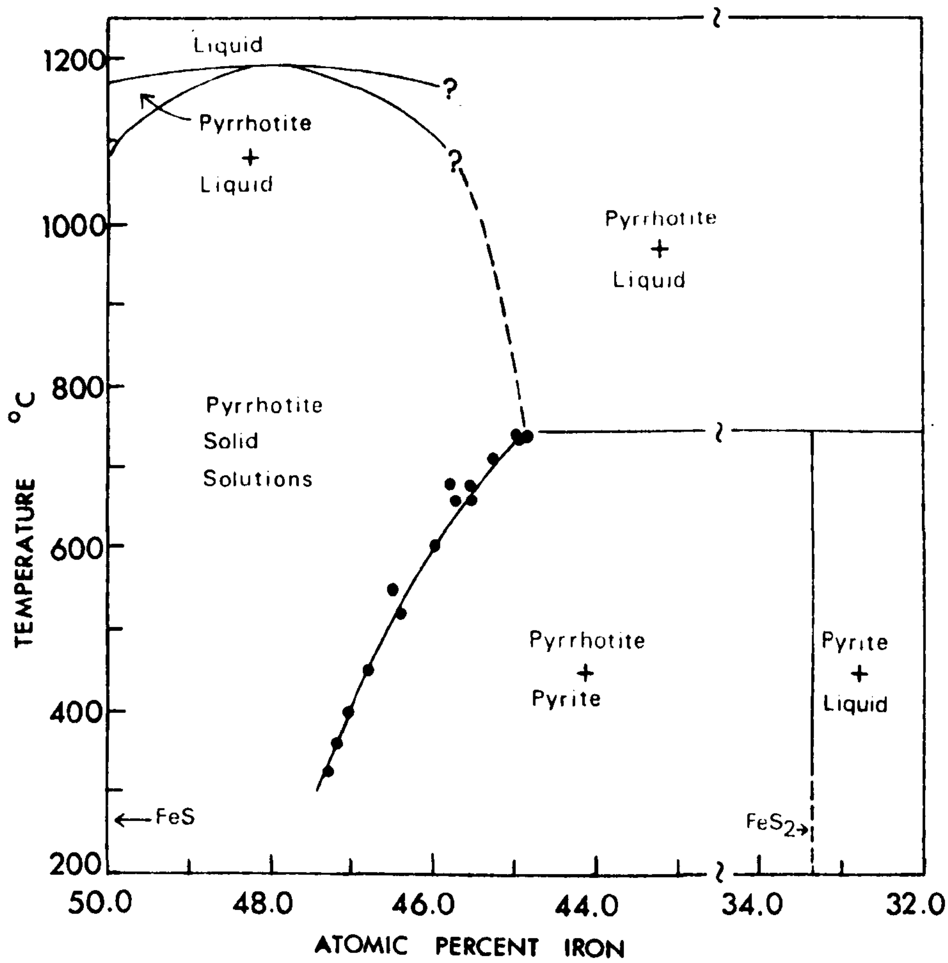


FIG. 10·1 Composition of pyrrhotite in equilibrium with pyrite versus temperature (from Arnold, 1962).

pyrrhotite deposits.

Toulmin and Barton (1964) confirmed the solvus of Arnold by experiment and projected it down to 200°C. They also linked the fugacity of sulphur to temperature and composition. Arnold had made sure that his assemblages lay well within the pyrite-pyrrhotite field, so that the pyrrhotite composition could not be due to sulphur deficiency in the system. Toulmin and Barton associated both sulphur fugacity (f_{S_2}) and pyrrhotite activity (a_{FeS}) as follows:

$$\log_{10} f_{S_2} = (70.03 - 85.83N)(1000/T - 1) + 39.30 \sqrt{1 - 0.9981N} - 11.91$$

and

$$\log_{10} a_{FeS} = 85.83(1000/T - 1)(1 - N + \ell n N) + 39.30 \sqrt{1 - 0.9981N} - 32.23 \tan h^{-1} \sqrt{1 - 0.9981N} - 0.002$$

where, N = mole fraction of FeS in pyrrhotite in the system

= 2 x atom fraction of Fe in pyrrhotite

and, T = absolute temperature.

The two quantities of sulphur fugacity and pyrrhotite activity are related by the Gibbs-Duhem equation:

$$N_{FeS}\, d.\log a_{FeS} = (NFeS - 1) d.\log f_{S_2}$$

Thus, by varying the activity of sulphur in a system it is possible, at a constant temperature (and pressure) to produce differing pyrrhotite compositions. Toulmin and Barton showed this graphically (Figs. 10.2 and 10.3). In the present case, both f_{S_2} and a_{FeS} can be fixed providing the real or experimental system is buffered by the presence of pyrite. The reaction:

$$FeS + \tfrac{1}{2}S_2 \rightleftharpoons FeS_2,$$

has an equilibrium constant, K, which is derived as:

$$K = \frac{aFeS_2}{aFeS.(aS_2)^{0.5}} \qquad \text{where} \quad aS_2 \approx fS_2 \ .$$

At any given temperature, the fugacity of sulphur and activity of pyrrhotite is fixed, as long as pyrrhotite and pyrite coexist in equilibrium. Addition of sulphur to the system will not change the fugacity, only the relative amounts of each phase. In a system where pyrite is not present, changes in sulphur fugacity result in changes in pyrrhotite composition to restore the original fugacity value at any given temperature.

Froese and Gunter (1976) calculated that there was no substantial effect of pressure on the solvus (Fig. 10.4). Thus the thermometer appeared to be workable down to about 200°C. However, Desborough and Carpenter (1965) cast doubt on the validity of the thermometer, since the pyrrhotite composition may not be unique at a given temperature, once the assemblage has cooled to below the beta-transformation line at around 315°C (Fig. 10.5). They agreed that down to this temperature, pyrrhotite composition in equilibrium with pyrite is a function of temperature. However, at about 315°C, the pyrrhotite reacts with the

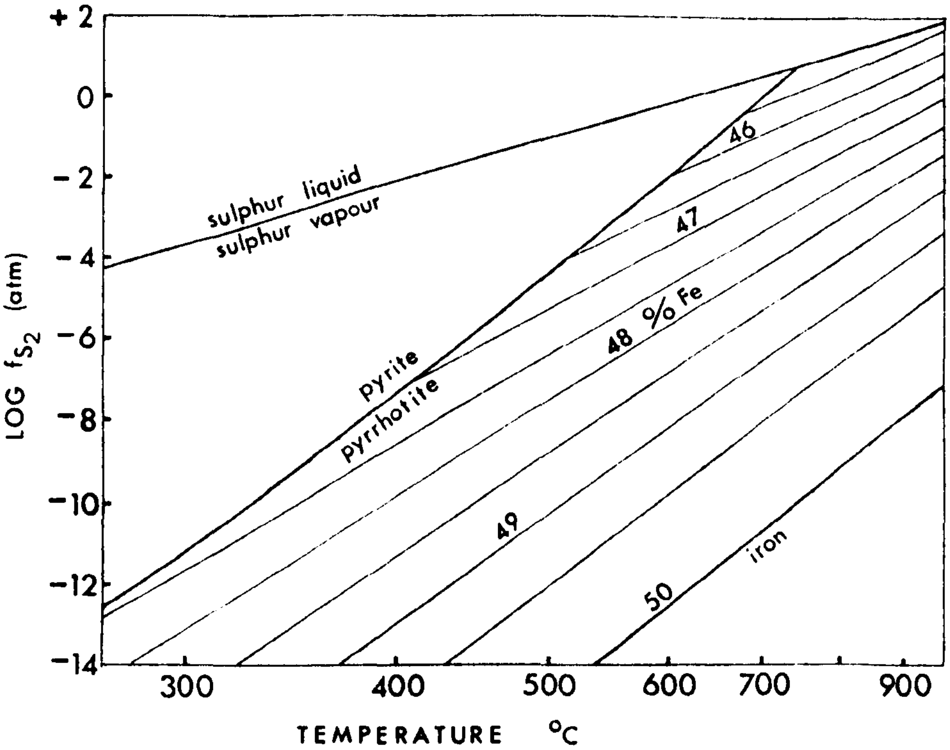


FIG. 10·2 Variation of pyrrhotite composition with temperature and sulphur fugacity (from Barton and Toulmin, 1964).

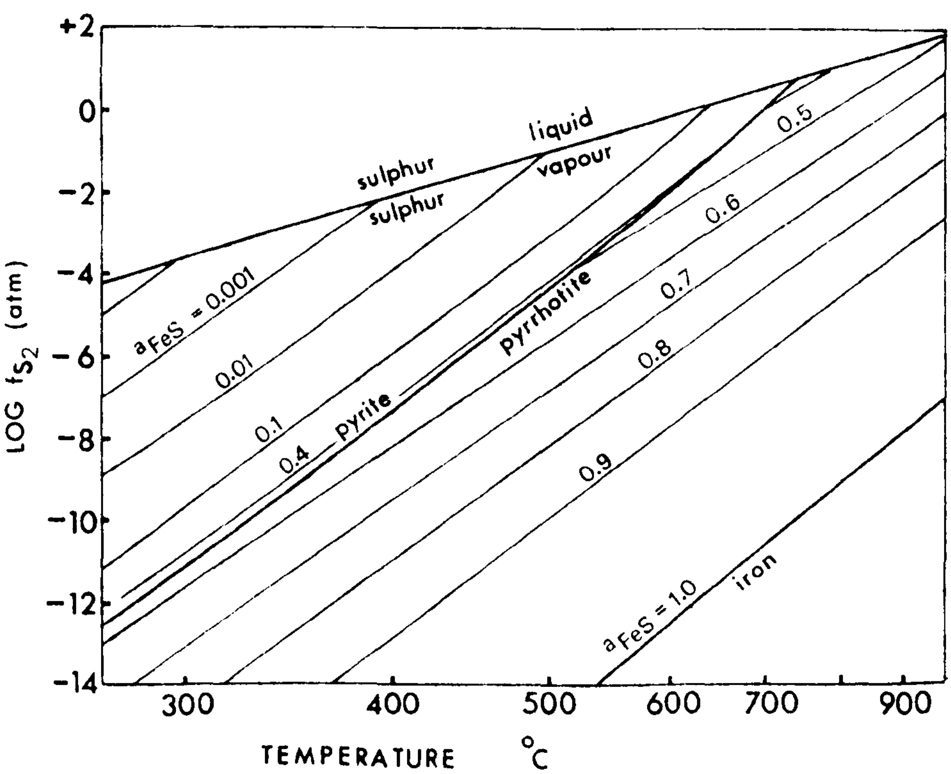


FIG. 10·3 Variation of pyrrhotite activity with temperature and sulphur fugacity (from Barton and Toulmin, 1964).

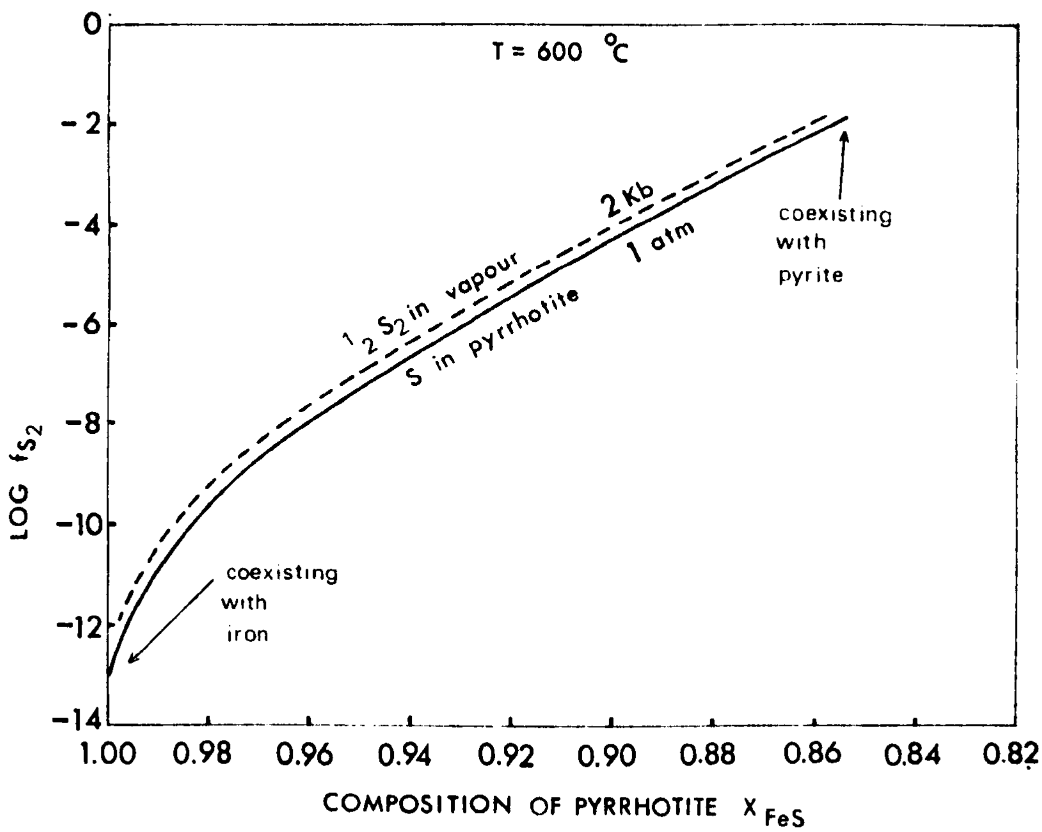


FIG. 10·4 Effect of pressure on the pyrrhotite - pyrite solvus (from Froese and Gunter, 1976).

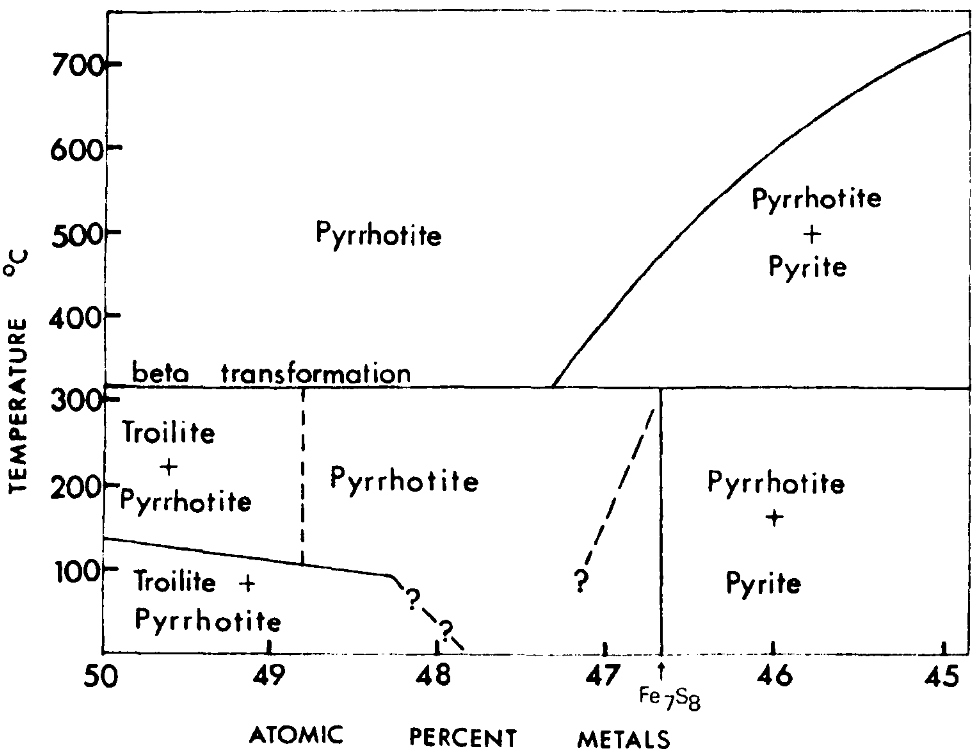


FIG. 10·5 Pyrrhotite – pyrite relationships down to 50 °C (from Desborough and Carpenter, 1965).

pyrite. According to their phase diagram, original compositions of pyrrhotite more iron rich than 46.67 atomic percent iron (Fe_7S_8) would, on cooling to the beta-transformation, react to form an iron rich phase ($\approx$ 48.6-49.0 atomic percent iron) and an iron deficient phase (Fe_7S_8). This low iron composition corresponds to a temperature of about 470°C on the solvus curve and thus temperature values may be seriously in error using this method. For this pyrrhotite geothermometer to be applicable, Desborough and Carpenter stated that rapid quenching would be necessary to 'freeze-in' the system. They consider that the equilibration is too rapid down to the beta transformation for quenching to occur. Original compositions less iron rich than Fe_7S_8 would, on cooling, produce pyrrhotite of this composition plus pyrite.

Pyrrhotites formed above the beta-transformation crystallize in the hexagonal system. At the beta-transformation, the resultant iron-rich pyrrhotite is hexagonal, while the iron-poor (Fe_7S_8) form is monoclinic. X-ray diffractograms of pyrrhotites from Grøsli and Eiker show that both hexagonal and monoclinic (iron rich and iron poor) forms are present and that the pyrrhotites have thus cooled below the beta-transformation line.

Desborough and Carpenter also state that oxidation can produce iron deficient pyrrhotites, e.g.

$$4Fe_9S_{10} + 3O_2 \rightleftharpoons 4Fe_7S_8 + 2Fe_2O_3 + 4FeS_2$$

and that this reaction can occur up to 600°C. They state that pyrrhotites more iron deficient than Fe_7S_8 do not occur in nature. However, such pyrrhotites do exist at Eiker. They may be of two possible origins.

Firstly, they may have been produced by an even greater degree of oxidation, possibly produced at a later stage in the ore body's history

$$(\text{e.g.}\quad 2Fe_7S_8 + O_2 \rightleftharpoons 2Fe_5S_6 + Fe_2O_3 + 2FeS_2).$$

Secondly, these iron poor pyrrhotites may represent metastable assemblages at, or below, the beta-transformation line, where the slow kinetics of the reaction prevented the formation of Fe_7S_8.

Sawkins et al (1964) studied iron-deficient pyrrhotites deposited by hydrothermal fluids and found large discrepancies between the thermometer values and fluid inclusion values. In connection with this, Desborough and Carpenter state that reaction of a more iron-rich pyrrhotite with sulphide rich solutions may well form iron-poor pyrrhotites. Limited hydrothermal remobilization of the Grøsli ores has occurred while oxidation of the ore deposit at Eiker is also evident. These localised processes plus the reactions occurring at the beta-transformation line indicate that the values derived from this method may be of limited significance.

10.3 THE SPHALERITE GEOTHERMOMETER

This thermometer is concerned with the iron content of sphalerite in equilibrium with pyrite and pyrrhotite. Early work by Kullerud (1953) on the FeS-ZnS solvus was in error because sulphur fugacity was not buffered in his experiments. Toulmin and Barton (1966) ran experiments using pyrrhotite sliding buffers to maintain the sulphur fugacity. The reaction of interest here is:

$$2FeS_2 \rightleftharpoons S_2 + 2FeS \quad \text{(in sphalerite).}$$

The iron content of the sphalerite is directly linked to the sulphur fugacity, at a given temperature and pressure. Toulmin and Barton represented this graphically (Fig. 10.6). They calculated experimentally that the iron content of sphalerite rises from about 13 mole percent FeS at 742°C to 19 mole percent at 580°C (Fig. 10.7). From Fig. 10.7 it is evident that in using this thermometer, sphalerite must be in equilibrium with both pyrrhotite and pyrite, to define the univariant solvus. Toulmin and Barton stated "most analyses of natural sphalerites, thought to have formed in equilibrium with pyrite and pyrrhotite well below 600°C show less than 20 mole percent FeS and there are some well-documented analyses with less than 12 mole percent. Thus we have strong reasons to suspect that the curve in fact does reverse its slope below about 600°C." This fact casts serious doubt on the use of the thermometer.

Subsequently, much experimental work was attempted to fix the solvus below about 580°C. Boorman (1967) overcame the difficulties of slow reaction rates below 580°C by using halide fluxing techniques. He stated that the solvus was vertical from 600° to 300°C with a fixed composition of about 21 mole percent FeS. Chernyshev and Anfilogov (1968) suggested that Boorman's experiments did not reach equilibrium and that the iron content (according to their experiments) increased down to 250°C where it had a value of 31 mole percent FeS. Einaudi (1968) concluded that despite many extrapolations below 600°C it was by no means certain which, if any, of the lines was correct and that the iron content of sphalerite was still dubious as a geothermometer.

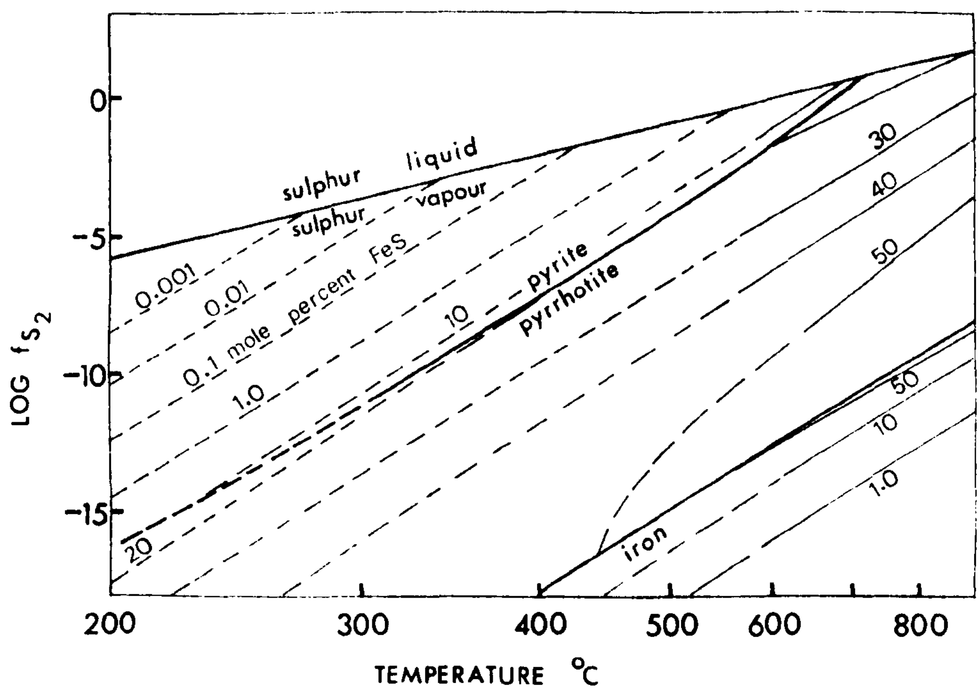


FIG. 10·6 Variation of iron content in sphalerite with temperature and sulphur fugacity (from Barton and Toulmin, 1966).

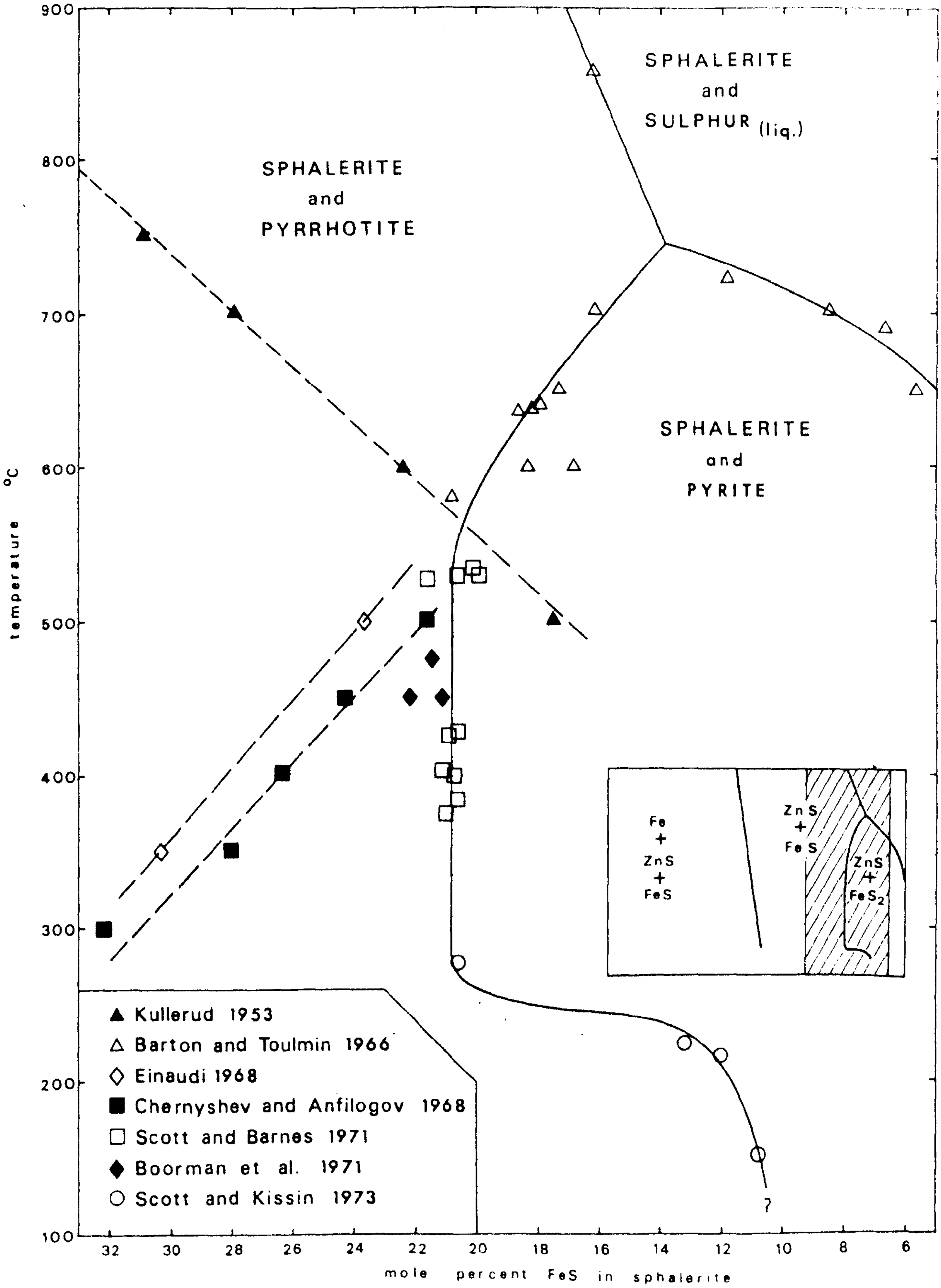


FIG. 10·7 Variation of iron content in sphalerite (in equilibrium with pyrite and pyrrhotite) with temperature

Campbell and Williams (1968) attempted to evaluate the experimental information by a detailed study of the sphalerites from the Quemont Mine, Quebec. They concluded that their data best fitted the curve of Toulmin and Barton (1966), but that the pyrite and sphalerite had ceased to react at lower temperatures. Thus they drew no conclusion about the lower part of the solvus.

Boorman et al (1971) used new analytical techniques to overcome disequilibrium at lower temperatures and stated that the FeS content of sphalerite is fixed below 550^{o}C at 20.8 mole percent. Scott and Barnes (1971) ran experiments down to 300^{o}C. They agreed with the results of Boorman et al and stated that the content is fixed at 20.7 $\pm$ 0.6 mole percent below 550^{o}C, thus making the thermometer useless below this temperature. Scott and Kissin (1973) suggested that below 265^{o}C the solvus ceases to be vertical and, in fact, becomes strongly negative (Fig. 10.7). They stated that the reason for the reversal of slope was because, at lower temperatures, the pyrrhotites of the assemblage have higher sulphur contents. This results in an increase in the sulphur fugacity, which causes a decrease in the iron content of the sphalerites. Groves et al (1974) concluded that use of the thermometer is rather speculative where low temperature re-equilibration may be possible. In fact, the reversal of the solvus makes the geothermometer inapplicable in most cases.

10.4 GEOBAROMETRY

During their early work on the sphalerite geothermometer, Toulmin and Barton (1966) also calculated that the solvus would be shifted to lower values with increasing pressure, since the presence of iron causes an increase in the volume of the sphalerite lattice. Scott and Barnes (1971) produced a graph of the effects of pressure on the iron content using calculations of theoretical pressure effects up to 6kb and actual experimental runs up to 1kb. They compared (with some success) their calculated pressures with those calculated from fluid inclusion studies for five natural occurrences. Scott (1973) experimentally located the solvus isobars for the assemblage sphalerite-pyrite-pyrrhotite between 325° and 710°C at 2.5, 5.0 and 7.5kb. He concluded that the geobarometer could be applied between the slope reversal in the solvus at 265°C/1 bar and the three-phase invariant point for the assemblage at 743°C/1 bar and 810°C/5kb. Lusk et al (1975) examined actual sphalerite compositions from the Noranda mine, Quebec and concluded that the geobarometer should be further investigated around 8kb in the range 300° to 500°C. Scott (1976) analysed sphalerites from three mines in regionally metamorphosed terraines and using this geombarometer obtained values up to 5.7kb, which were in agreement with other pressure estimates. Lusk and Ford (1978) improved the geobarometer in the high pressure region and stated that it could be applied at least up to 10kb (Fig. 10.8).

Scott and Barnes (1971) warned that sphalerites which have been metamorphosed below 525°C are heterogeneous in their iron content and contain iron enriched patches. This heterogeneity is strongly dependent on temperature. They state that the patches may represent meta-stable high temperature polytypes of sphalerite. They produced a graph (Fig. 10.9) showing temperature versus difference in iron content between patches and host and stated that this may prove to be a useful

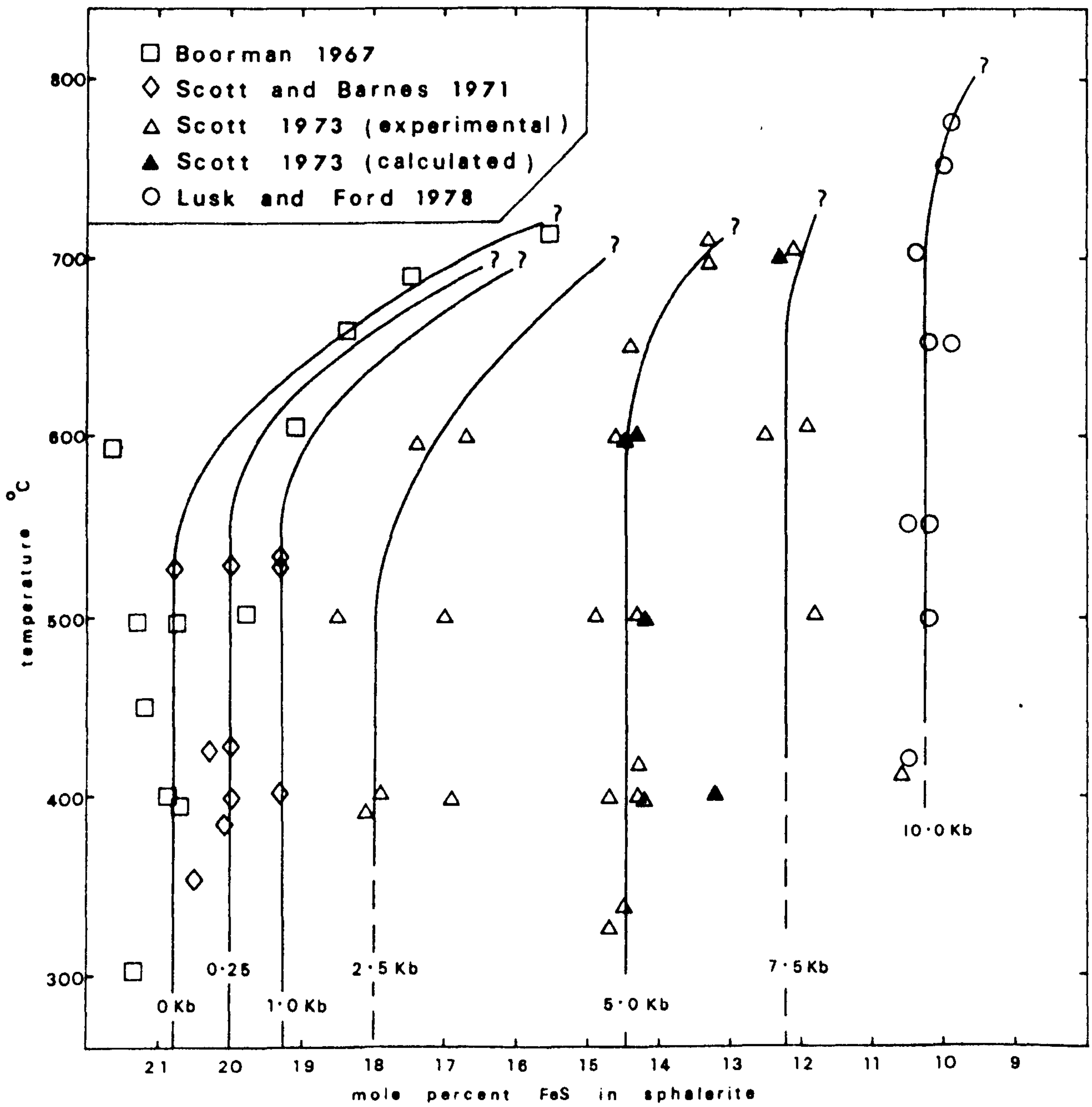


FIG. 10·8 Variation of iron content in sphalerite as a function of pressure and temperature

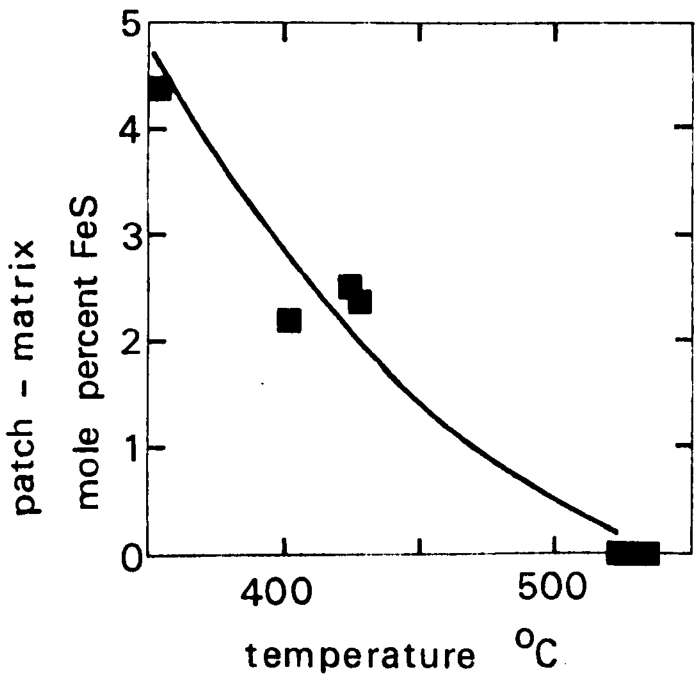


FIG.10·9 Relationship between temperature and difference in iron content between patches and matrix of sphalerite (from Scott and Barnes, 1971) thermometer, although much work still needed to be done.

10.5 APPLICATION OF THE METHODS TO THE ORES AT GRØSLI AND EIKER

The use of the pyrite-pyrrhotite geothermometer (Arnold, 1962) is highly suspect. However, temperatures derived by using this method and by using the 'iron-rich patch' method of Scott and Barnes (1971) are presented in Tables 10.3 and 10.4. Pressure values are presented in Table 10.5. The supracrustal country rocks have undergone two phases of amphibolite facies metamorphism and the ores have also been subjected to at least one of these events and probably underwent the earlier Svecofennian event also. Thus it is reasonable to assume that all three phases were in equilibrium before cooling associated with the last metamorphism. Pyrrhotite probably re-equilibrates very rapidly to lower temperatures, but Scott (1973) was of the opinion that unless sphalerite is deposited hydrothermally or heated to very high temperatures (possibly over 600°C) then it will not be in equilibrium with the other mineral phases. However, once equilibrated the kinetics of iron diffusion in the solid state inhibits a change in composition with cooling. Possibly on heating, during metamorphism, the sphalerite developed higher temperature patches, since the temperature was not high enough for total re-equilibration: these patches, not re-equilibrated on cooling thus recorded the temperature of metamorphism.

TABLE 10.3

FORMATION TEMPERATURE AS CALCULATED FROM THE PYRRHOTITE-PYRITE SOLVUS (^{o}C)

ANALYSIS	TEMP.	AVERAGE	RANGE (2σ)
1	402		
2	343	397	360 - 435
3	418		
4	426		
5	426		
6	430		
7	428	432	412 - 452
8	412		
9	466		
10	BELOW 315		
11	322		
12	BELOW 315		
13	BELOW 315		
14	BELOW 315		
15	435		
16	489	462	435 - 489
17	462		
18	520		
19	509	516	510 - 523
20	520		
21	582		
22	636	641	598 - 684
23	677		
24	668		

TABLE 10.3 continued

1 to 4	Grøsli.	Borehole 7.	Epidote amphibolite.
5 to 9	Grøsli.	Borehole 11.	Gneiss with hydrothermal alteration.
10 to 14	Grøsli.	Borehole 4.	Gneiss with much hydrothermal alteration (gangue).
15 to 17	Grøsli.	Borehole 10.	Metagabbro.
18 to 20	Grøsli.	Borehole 6.	Gneiss with no hydrothermal alteration.
21 to 24	Eiker.		Massive sulphide ore.

TABLE 10.4

EQUILIBRATION TEMPERATURE AS CALCULATED FROM IRON PATCHES IN SPHALERITE (^{o}C)

ANALYSIS	AVERAGE	RANGE (2σ)
1	461	450 - 484
2	427	400 - 458
3	473	425 - 535
4	430	389 - 491
5	453	424 - 496
6	473	446 - 513

1	Grøsli.	Borehole 7.	Epidote amphibolite.
2	Grøsli.	Borehole 11.	Gneiss with hydrothermal alteration.
3	Grøsli.	Borehole 4.	Gneiss with much hydrothermal alteration (gangue).
4	Grøsli.	Borehole 10.	Metagabbro.
5	Grøsli.	Borehole 6.	Gneiss with no hydrothermal alteration.
6	Eiker.		Massive sulphide ore.

TABLE 10.5

EQUILIBRATION PRESSURE AS CALCULATED FROM SPHALERITE (kb)

ANALYSIS	PRESSURE	AVERAGE	RANGE (2σ)
1	7.0		
2	6.8		
3	5.6	6.4	5.7 - 7.1
4	5.6		
5	6.9		
6	4.8		
7	6.8	6.4	5.3 - 7.5
8	6.6		
9	7.4		
10	6.3		
11	7.4		
12	7.0	7.1	6.4 - 7.8
13	8.0		
14	6.6		
15	6.3		
16	4.4		
17	5.0	5.8	4.8 - 6.8
18	6.5		
19	6.7		
20	5.5		
21	6.3	5.5	4.8 - 6.2
22	5.7		
23	4.6		
24	8.4		
25	7.6		
26	8.3	7.8	7.2 - 8.4
27	7.9		
28	7.0		

TABLE 10.5 continued

1 to 5	Grøsli.	Borehole 7.	Epidote amphibolite.
6 to 9	Grøsli.	Borehole 11.	Gneiss with hydrothermal alteration.
10 to 14	Grøsli.	Borehole 4.	Gneiss with much hydrothermal alteration.
15 to 19	Grøsli.	Borehole 10.	Metagabbro.
20 to 23	Grøsli.	Borehole 6.	Gneiss with no hydrothermal alteration.
24 to 28	Eiker.		Massive sulphide ore.

10.6 TEMPERATURE-PRESSURE VALUES AT GRØSLI AND EIKER

Formation temperatures, derived by the pyrite-pyrrhotite solvus method, range from over 500°C to below 315°C, for the Grøsli deposit. Temperatures using the sphalerite patches all gave 450°C $\pm$ 20°C. Pressure calculations range from 5.5 to 7.1kb. These temperatures and pressures indicate a lower epidote-amphibolite grade equilibration of the ores. The five borehole samples analysed contain ores with differing histories. The ore sample from borehole 7 is contained in a low grade epidote-amphibolite, while that from borehole 10 is in a metagabbro. The specimens from boreholes 6 and 11 are in gneisses and that from borehole 4 is in gneiss replaced by gangue and ore minerals. The difference in the two temperatures obtained (by solvus and iron-rich patch methods) for the sample from borehole 7 can be ascribed to the fact that, in the case of the pyrite-pyrrhotite geothermometer, partial equilibration of the pyrrhotite towards the Fe_7S_8 sub-beta transformation composition has occurred; slow reaction kinetics produced the present metastable composition which does not refer to the pyrrhotite-pyrite solvus position but to an equivalent composition along the beta-transformation line (Fig. 10.5). The same explanation can be applied to the other borehole specimens.

The sample from borehole 4 contains ore material which has been highly remobilized by hydrothermal fluids, while the sample from borehole 6 contains ore with no evidence of such processes. This fact may indicate that the ore in borehole 6 is in the least altered state. Since the specimen from borehole 4 is within the most hydrothermally altered host rock of the group, the ore may be expected to exhibit the most altered features. It is interesting in this context that, relative to all other sample, the 'least altered' specimen contains pyrrhotites with the lowest iron contents and sphalerites with the highest iron values. Conversely, the 'most altered' specimen contains pyrrhotites with the highest iron content and sphalerites with the lowest iron values. It would appear therefore, that the iron has redistributed itself during the hydrothermal fluxing of the ores.

The samples from the Eiker mine indicate temperatures of the epidote amphibolite facies but higher pressures than those of Grøsli. The large discrepancy between the pyrite-pyrrhotite temperature determinations and the much lower values of the iron-rich patch method, is explained by the fact that much oxidation of the ore body has occurred, and produced iron-poor pyrrhotites.

10.7 CONCLUSIONS

Due to the factors mentioned above, temperature calculations from the pyrite-pyrrhotite solvus must be considered of limited significance. Temperatures deduced from the iron-rich patches in the sphalerites must probably be treated as estimates. The method, of Scott and Barnes (1971), needs careful study and more experimental work to confirm their findings. One important factor is that their graph is drawn from data for experiments up to 1kb only. The development of the iron-rich patches may be inhibited at higher pressure because of increased lattice size with increased iron content. Therefore, the patch-matrix differences may be smaller than at 1 kb, thus making the calculated temperature higher. This fact may explain results from the hydrothermally re-deposited ores at Grøsli, which were introduced after the epidote-amphibolite overprint and are greenschist facies mobilizations (as shown by the gangue minerals) but show temperatures in the epidote-amphibolite facies.

Pressure calculations must also be viewed with caution because of lack of equilibrium in the sphalerite crystals, but the Grøsli deposit appears to have been metamorphosed in the range of 5.5 to 7.0kb while the Eiker ores have been affected at a slightly higher pressure of about 8kb. Even with very accurate analytical results, the slopes of curves used in the determinations are steep enough to produce substantial variations in estimates. This factor is shown up in the ranges of temperature and pressure in Tables 10.3 to 10.5.

Despite the approximate nature of the results, they do indicate that the last event to affect the ores was a medium to low grade regional metamorphism. Fig. 10.10 shows the range of pressures and temperatures affecting the two mine areas. The Grøsli deposit has a range which is bisected by the 20°C/km geotherm which is considered as the average gradient for medium pressure regional metamorphism (Miyashiro, 1973) and corresponds to a depth of 20-25km. The Eiker deposit was metamorphosed in a slightly higher pressure regime between the 20°C/km and 15°C/km gradients and corresponding to a depth of 25-30km.

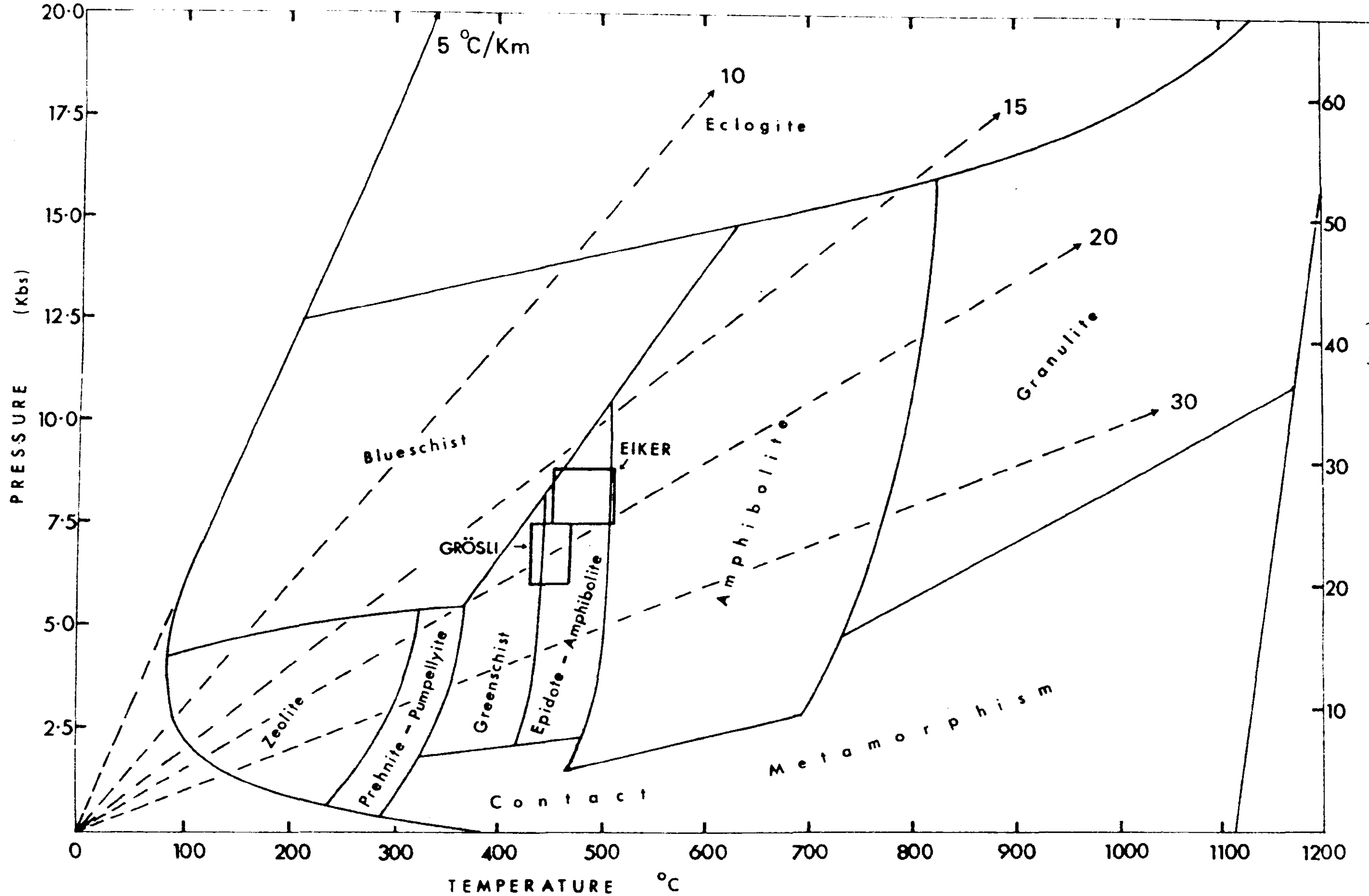


FIG. 10·10 Ranges of pressures and temperatures of metamorphism which affected the Grösli and Eiker deposits.

CHAPTER ELEVEN

THE ORIGINS OF THE ORE DEPOSITS

11.1 INTRODUCTION

The Precambrian supracrustals of the Kongsberg Series yield maximum radiometric dates of 1600 m.y.B.P. Rocks of this age have been involved in two regional metamorphisms at Middle to Upper Amphibolite facies grade, while microscopic evidence also strongly suggests that the ore deposits at Grøsli and Eiker have been involved in at least one and probably both of these events. The genesis of the ore body at Grøsli proves relatively less of a problem than that at Eiker in view of the evidence presented. Since the host rocks are so completely metamorphosed, their original characteristics (and hence the environments of deposition of the ore bodies) are difficult to determine. However, from the evidence given in the previous chapters, several environments and modes of deposition can be examined and discussed.

Four possible origins are given below and will be examined in turn. The development of the ore bodies can be ascribed to both original and metamorphic features and various processes have contributed to their present form.

The four possible origins are:

1. An association with the basic intrusives.
2. A syn-volcanic origin.
3. A hydrothermal injection and/or replacement origin.
4. A sedimentary depositional origin.

11.2 AN ASSOCIATION WITH THE BASIC INTRUSIVES

At both Grøsli and Eiker the sulphides are closely associated with the basic intrusives. At Grøsli the ore could be envisaged as a magmatic differentiate (since it often occupies interstices between crystals in the gabbro), or as a late stage deposit (since sulphides sometimes vein the basic rock and are associated with heavy chloritization). At Eiker relationships are less clear because of the intense shearing of the ore deposit, but the occurrence of the ore could be seen as an accumulation in the amphibolite body.

Basic rock-sulphide associations are world-wide and embrace some of the largest known deposits, including Sudbury (Ontario, Canada), Lynn Lake (Manitoba, Canada), Noril'sk (Siberia, U.S.S.R.) and Kotalahti (Sova, Finland). In Norway the nearest sulphides associated with coarse grained basic intrusives occur in the Ringerike and Bamble areas (see Fig. 11.1).

Two very important differences exist between the ore deposits of Grøsli and Eiker and those given above. First, the basic rocks of the Ringerike and Bamble areas tend to be highly mafic with isolated developments of ultra-mafics. The intrusives at Grøsli and Eiker do not show any tendency towards ultra-mafic compositions, although the serpentinite pod present at Eiker may be indication of such rocks at depth. Second, and most important, is the concentration of nickel in the ore deposits. Stanton (1972) divides basic magmatic ore deposits into two groups; a chromium-nickel-platinoid association and an iron-nickel-copper sulphide-platinoid association. The former group is essentially an oxide mineralization and not relevant to the present discussion. Table 11.1 shows the Ni/Cu ratio for the Grøsli and Eiker

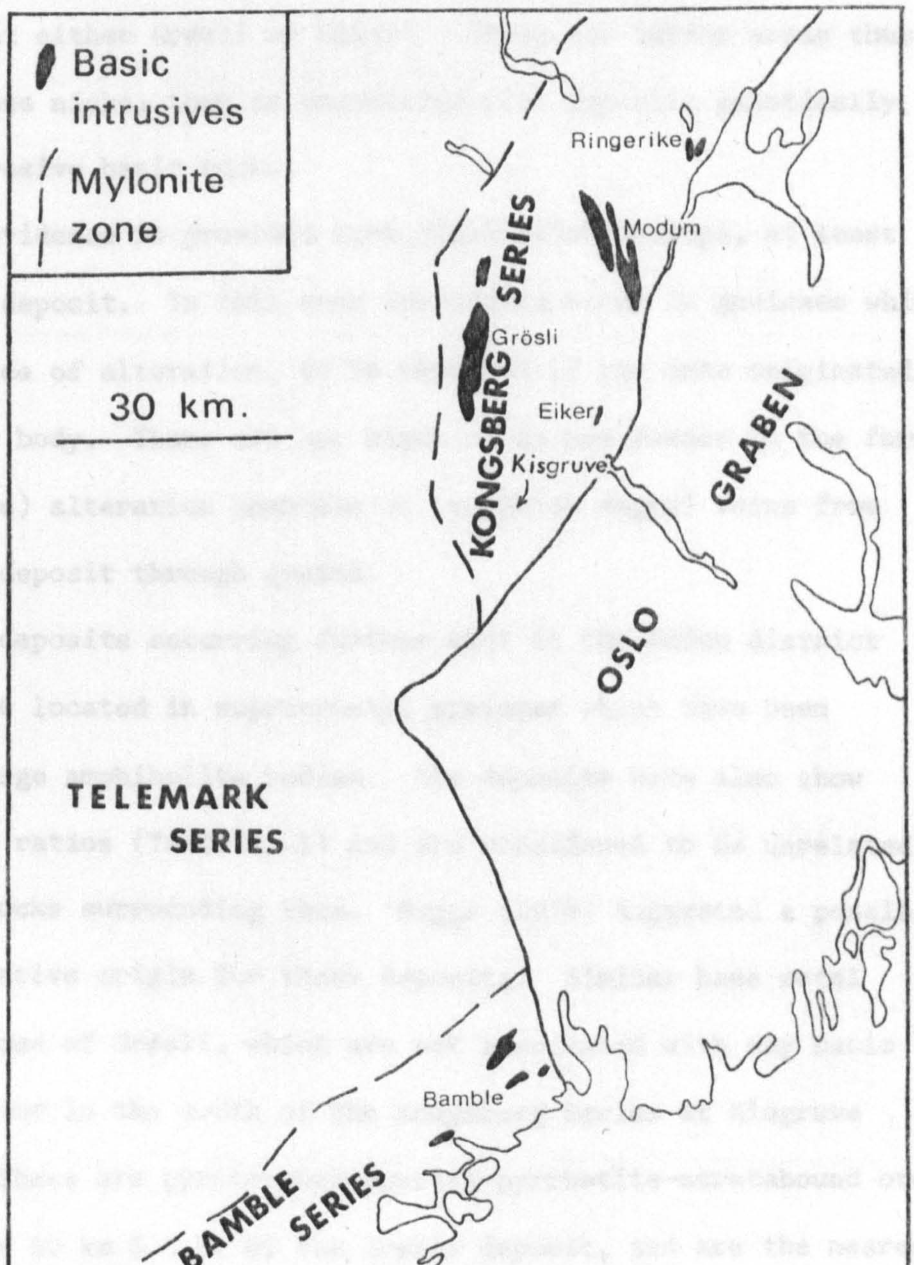


FIG.11.1 Sulphide deposits associated with basic intrusives.

areas and for the examples above which commonly contain pentlandite (not present at either Grøsli or Eiker). These two latter areas thus contain far less nickel than is associated with deposits genetically linked to intrusive basic rocks.

Further evidence is provided from field relationships, at least in the Grøsli deposit. In this area ore bodies occur in gneisses which show no evidence of alteration, to be expected if the ores originated from the basic body. There are no signs of an ore-feeder in the form of (fluid-borne) alteration channels or (sulphide magma) veins from gabbro to ore deposit through gneiss.

Sulphide deposits occurring further east in the Modum district (Fig. 11.1) are located in supracrustal gneisses which have been intruded by large amphibolite bodies. The deposits here also show very low Ni/Cu ratios (Table 11.1) and are considered to be unrelated to the basic rocks surrounding them. Bugge (1978) suggested a possible volcanic exhalative origin for these deposits. Similar base metal deposits to those of Grøsli, which are not associated with any basic intrusives, occur in the south of the Kongsberg Series at Kisgruve (Fig. 11.1). These are pyrite-chalcopyrite-pyrrhotite-stratabound ores occurring about 20 km S.S.E. of the Grøsli deposit, and are the nearest deposits of their type to Grøsli. However, any suggestion that these two deposits may represent lateral equivalents is highly speculative.

11.3 A SYN-VOLCANIC ORIGIN

Both ore deposits could have been strata-bound Fe-Cu-Zn sulphides which have many features in common with the volcanic exhalative or

TABLE 11.1

Ni/Cu RATIOS OF BASIC ASSOCIATED ORES

	Deposit	% Ni	% Cu	Ni/Cu
1	Sudbury	1.62	1.29	1.3 :1
2	Lynn Lake			
3	"A plug"	1.20	0.60	2 :1
4	"EL plug"	4.50	1.50	3 :1
5	Noril'sk	0.40	1.00	0.4 :1
6	Kotalahti	0.80	0.30	3.8 :1
7	Bamble	av. 2.75	av. 0.75	av. 3.7 :1
8	Ringerike			
9	Grøsli	0.03	0.36	0.08:1
10	Eiker	0.005	0.06	0.08:1
11	Modum	0.02	0.15	0.10:1

Analyses from Grøsli and Eiker (9 and 10) represent the highest Ni/Cu ratios of those two deposits.

Anal. 1-6. From Stanton (1972).

Anal. 7, 8, 11. From Bugge (1978).

volcanic marine stratiform and stratabound deposits recognized world-wide. The genetic models of syn-volcanogenic ore deposits are diverse, involving differing host rocks deposited in differing environments in response to distinct plate tectonic settings and various stages in their evolution. A resume of the history of these theories and a definition of the terms used are thus first discussed.

11.3.1 Resume of previous work on syn-volcanic massive sulphides

Earlier workers, especially De Baumont (1847), Carstens (1919) and Ohashi (1919) had noticed the link between stratiform sulphides and volcanic activity, but in the late 1950's theories of sedimentary, volcano-sedimentary and exhalative-sedimentary ore genesis began to gain increasing popularity. King et al (1953) were among the first to indicate these processes by suggesting the Broken Hill deposits of New South Wales to be sedimentary. Stanton (1955) further proposed a link between iron sulphide stratiform ores and volcanic-arc regimes, while Knight (1957) drew attention to the fact that many major ore deposits were spatially unrelated to any sizeable plutonic intrusions (necessary for magmatic or hydrothermal alternatives).

Papers published by workers such as Oftedahl (1958), Williams (1960) and Kinkel (1962) further supported and strengthened the theories. In the following years much work was concerned with the actual modes of deposition and emplacement of these ores (e.g. Borchert, 1960; Vokes, 1966; Kinkel, 1966; Smirnov, 1968 and Anderson, 1969).

Stanton (1972) summarized the ideas on origins of the deposits and defined two principal theories. The first theory was that the deposits were due to subsurface hydrothermal replacement of pre-existing rocks. Stanton stated that variations on this major theme could be produced by examining the possible alternative origins of the metals and sulphur. The deposits could have been derived by replacement of host rock by both metal and sulphur from either plutonic intrusions or from leaching and lateral secretion of host rocks. Also, it is possible that pre-existing sedimentary iron sulphides were substituted by metals from intrusions or leaching. Finally, sulphurous volcanic gas could have fixed the metals within host rocks and thus formed sulphides.

The second principal theory defined by Stanton was that the sulphides were sub-aqueous sedimentary deposits added contemporaneously to the host rocks. Variations on this theme again involve metal and sulphur derivation. The two may be derived entirely from sea water, the sulphur being reduced from sulphate by bacteria. The metals may also be derived from continental erosion. Both sulphur and metals may have been derived from volcanic hot springs or deep geothermal brines. Finally, sulphide may be formed within volcanic vents and emitted as suspensions, tuffs and breccias.

The two alternative theories of formation are often very difficult to apply to given deposits, especially where an ore has been metamorphosed, but Stanton favours the idea that most major ore deposits are either volcano-sedimentary or sedimentary in origin and not formed by late stage replacement of older host rocks.

Vokes and Gale (1976) defined the deposits which have a direct genetic connection with volcanism as 'synvolcanic' or 'volcanogene'. They considered that the ore deposits may have formed by one of two processes, either deposited from fluids derived from the volcanic magma or deposited from connate and meteoric waters which leached host rocks. Recent research has involved the identification of the origin of ore forming solutions using stable isotopes (e.g. Sheppard, 1977), the study of geothermal models to elucidate cycles of water movement associated with volcanic mineralization (Henley and Thornley, 1979) and the behaviour of hot saline solutions on extrusion onto the sea floor (Turner and Gustafson, 1978).

In the late 1960's interest turned to the tectonic environments associated with ore deposits. Sillitoe (1973) criticised earlier workers such as Gilmour (1971), Hutchinson and Hodder (1972) and Jenks (1971) for "application of imprecise geosynclinal terminology" to describe environments of ore formation. He stated that the deposits could be described in terms of constructive and destructive plate margins. Since the early 1970's workers have attempted to not only identify the plate-tectonic environment of particular ore deposits, but also to classify groups of deposits, according to ore-host rock environments, petrology and chemistry and to relate these to specific plate-tectonic locations. In the identification of environments of individual ore deposits the two main considerations are the mineralogy/chemistry of the ore deposit and the lithology/chemistry of the associated host rocks. In the latter case, the trace element chemistry of associated extrusives and intrusives (as developeed by Pearce and Cann, 1971, 1973) has proved very helpful in elucidating plate-tectonic regimes.

Classification of ore deposits has proved more difficult as more deposits are becoming recognized as volcanogenic. Stanton (1972) was one of the first to notice distinct environmental associations and his classifications, although somewhat subjective, used many of the characteristics which have been used by later workers (Table 11.2). He defined four groups, which merged into each other, according to whether the associated host rocks were volcanic, sedimentary or combinations of both. Where sedimentary rocks alone are present, Stanton stated that their depositional environments can vary tremendously, but are always of near-shore character. In his concluding statement he tended to the view that most of the cited deposits were tied to one phase of volcanism or another.

Hutchinson (1973) (Table 11.3) used the major element chemistry of ores to group the deposits into one of three categories associated with distinct tectonic environments. Sawkins (1976) (Table 11.4) considered that the petrochemistry and lithology of host rocks were of much more importance and criticised Hutchinson's classification for lack of consistency in the occurrence of Pb, especially in Precambrian terrains. Sawkins' classification is very much in agreement with Stanton's in advocating a near-shore environment for the deposition of associated sediments.

Vokes and Gale (1976) however, were sceptical of classifications such as those above. They considered that the knowledge of the true characteristics of 'type' ore deposits was still incomplete and that many of the so-called 'type deposits' might turn out to be atypical. They stated that as far as Scandinavian Caledonide ore deposits were concerned there were similarities to several 'type-deposits' (e.g. Kuroko, Besshi and Cyprus) but no complete analogy with any of them. A further consideration relevant to such classifications is that features of given tectonic environments may have changed markedly

TABLE 11.2. (after Stanton (1972). Simplified)

TYPE	HOST ROCKS	EXAMPLES	ENVIRONMENT
Sulphides spatially related to sedimentary and volcanic features.	Pyroclastics: basalt to rhyolite. Andesites/Dacites most common. Limestones, dolomites, silts, shale.	MacArthur R. Mt. Isa	Fairly shallow water, near shore. Substantial biological activity.
Sulphides spatially related to volcanism but no specific sedimentary features.	Pyroclastics: andesite to rhyolite. Calc-alkaline. Preferences usually for acid members, except Cyprus.	Cyprus, Superior, Kuroko.	Sub-aqueous marine or lacustrine - lagoonal.
Sulphides spatially related to specific sedimentary features but no volcanism.	Various: e.g. shales, bituminous, calc, dolomitic. argillites, conglomerates, etc.	Mansfeld Roan	- shallow, putrid near shore basins - well flushed neritic or sub-littoral.
Sulphides spatially related to no specific sedimentary or volcanic features.	Various: e.g. quartzites, calcareous argillites, siltstones, conglomerates, etc.	Sullivan, Broken Hill.	Shallow water. Possibly geosynclinal.

TABLE 11.3. (after Hutchinson (1973). Simplified)

BASE METAL TYPE	ASSOCIATED VOLCANIC ROCK TYPES	TYPE OF VOLCANISM AND SEDIMENTATION	TECTONISM	AGE
Zn - Cu - PYRITE	Basalt to rhyolite. Intermediate bulk composition (?) Tholeiitic and calc-alkaline.	Deep sub-aqueous building to shallow volcanic domes. Chemical precipitates. Immature volcaniclastics.	Early eugeo-synclinal. Major subsidence.	Archaean Proterozoic
			Early subduction	Pre-Ordovician Mid-Devonian
Pb - Zn - Cu - PYRITE	Andesite to rhyolite intermediate to felsic bulk composition calc-alkaline	Sub-aqueous to sub-aereal explosive centres. Minor chemical ppts. Greywackes, shales, volcaniclastics common.	Later eugeo-synclinal infilling and uplift	Proterozoic
			Later subduction	Ordovician, Triassic, Tertiary
Cu - PYRITE	Ophiolite suites. poorly differentiated	Deep sub-aqueous fissure eruptions. Chemical precipitates. Minor clastics.	Early stage of plate rifting.	Ordovician Cretaceous Jura-Cretaceous Cret-Eocene

TABLE 11.3 contd.

BASE METAL TYPE	AGE	EXAMPLES
Zn-Cu-	Archaean	Timmins, Ontario; Noranda, Quebec.
PYRITE	Proterozoic	United Verde, Arizona.
	Pre-Ordovician	Rambler, Newfoundland.
	Mid-Devonian	West Shasta, California.
Pb-Zn-	Proterozoic	Mt. Isa, Queensland; Errington and Vermillion,
Cu-	Ordovician	Sudbury, Ontario.
PYRITE	Triassic	Bathurst, New Brunswick.
	Tertiary	East Shasta, California.
		Kuroko, Japan.
Cu-	Ordovician	West Newfoundland.
PYRITE	Cretaceous	Cyprus.
	Jura-Cretaceous	Island Mtn., California.
	Cret-Eocene	Phillipines.

TABLE 11.4. (after Sawkins (1976). Simplified)

TYPE	HOST ROCKS	ENVIRONMENT	FEATURES	EXAMPLES
Kuroko	Felsic, calc-alkaline pyroclastics. Fe-Cu-Zn Fe-Cu-Zn-Pb	Submarine. Convergent plate margins.	Phanerozoic deposits contain more Pb. and sulphate gangue.	Noranda, Skellefte, Jerome, Bathurst, Buchans, Bleikvassli, Kuroko, E. Shasta.
Cyprus	Basic volcanics. Ophiolite complexes. Fe-Cu Fe-Cu-(Zn)	Submarine. Sea-floor or back-arc spreading.	Cu always dominates over Zn.	Troodos, Betts Cove, Halls Bay, Kure.
Besshi	Mafic volcanics. Thick greywackes. Fe-Cu Fe-Cu-Zn	Submarine. Arc-trench gap. Early arc convergence.	Appear to be related to early geosynclinal stages.	Sanbagawa, Abukuma, Birtavarre, Folldal, Rødhammer, Killingdal, Kvikne, Elizabeth Mine.
Sullivan	Thick clastic sequences. Cu-Fe, Cu-Fe-Zn, Cu-Fe-Zn-Pb.	Continental margins or intracontinental troughs associated with triple junctions.		Sullivan, Mt. Isa, Rammelsberg, Ducktown.

with time. This appears to be especially true in any comparison of Precambrian ores with those of the Phanerozoic.

More recently Garson and Mitchell (1977) reviewed the types of mineralization occurring at destructive plate boundaries and indicated relationships between the ore types in terms of their relative time and spatial positions within mobile belts. Instead of stating precise chemical characteristics of the ore deposit and defining groups based on one "type" locality, they emphasised the broad tectonic environments and noted features of ancient deposits which appeared similar to those exhibited by present day examples. Of particular relevance to the present discussion (of the Grøsli and Eiker deposits) they cited the Kuroko deposits of Japan as island-arc mineralizations, and noted the wide variability of tectonic setting and chemistry.

11.3.2 Possible environment of the Grøsli deposit

The ores at Grøsli are less problematical than those at Eiker. A genetic affinity to the Vinor bodies in the area appears very unlikely. The supracrustals could be mainly acid volcanics with a few inter-banded sediments. The most suitable grouping according to Stanton's divisions (Table 11.2) is that of an association with volcanism but with none of the special features of sedimentation. The enclsoing rocks are also acidic and mostly calc-alkaline, both of which are in accordance with the described features. Table 11.4 indicates that the affinities may be with the Kuroko group of deposits except for the high Co/Ni ratios of the Grøsli rocks. The deposit is of the Zn-Cu-pyrite grouping of Table 11.3 (after Hutchinson, 1973) and as such does not correspond to the Kuroko environment, as suggested in the previous two classifications. However, the parameters used to distinguish between this group and the group of deposits which include the Kuroko-type ores are open to criticism.

The primary difference between the two groups is the presence of Pb within the Kuroko type deposits. However, this element is known to be highly variable within these ores. Hutchinson states that Pb is low within Archaean deposits but higher in Proterozoic-Phanerozoic ores, while Sawkins (1976) stated that Pb was absent in Archaean ores and low in Proterozoic ores. Lambert and Sato (1974) note that a low Pb level is not uncommon in younger ores including some Miocence Japanese (Kuroko) deposits. With respect to the area under discussion, Hutchinson's two groupings (Pb-Cu-Zn-pyrite and Cu-Zn-pyrite) provide few criteria for discrimination.

11.3.3 Characteristics of Kuroko-type mineralization

The Grøsli deposits could be related to convergent plate margin environments or their Precambrian equivalents. The characteristics of Kuroko ores and older deposits of similar affinity will therefore be summarized. There is much published work on the Japanese deposits concerned with individual localities and all the deposits as a group. Reviews of their features have been produced by Tatsumi et al (1970), Matsukuma and Horikoshi (1970), Lambert and Sato (1974) and Sato (1977). The deposits are stratiform bodies associated with underlying stockwork ores and fissure-filling veins and are of Miocene age. The ore deposits are associated with waning phases of volcanic activity and with felsic tuffs, lavas and shallow intrusives, formed sub-aqueously in shallow marine environments, probably 100-200 m deep. The associated igneous rocks are of calc-alkaline affinity and rhyolitic in composition (Tatsumi and Clark, 1972).

Fig. 11.2 is a schematic representation of the formation of a 'typical' ore body. Tatsumi and Watanabe (1971) considered the basal stockwork ore and vent breccia ore to be formed epigenetically by the movement of hydrothermal fluids up through the lava domes and tuffs. The bedded deposits formed syngenetically by precipitation where ore fluids emanated at the submarine surface. The formation of the ore bodies appears to be linked to the progressive differentiation of an intermediate composition magma. Initial volcanism is usually of andesitic affinity, passing through dacitic to rhyolitic types. The "white rhyolite" domes are the most highly differentiated of the sequence. Mineralization appears to be very rapid and restricted to narrow stratigraphic horizons.

The ore deposits themselves are stratified and are characterized by their mineralogy and position in the sequence. 'Keiko' is the pyrite-chalcopyrite-quartz stockwork ore in which much of the original volcanic rock has been replaced by hydrothermally deposited silica. 'Oko' is a pyrite-chalcopyrite ore with minor sphalerite, barite and quartz: it usually occurs below the dominant 'Kuroko' body which is of sphalerite-galena-chalcopyrite-barite-pyrite composition and forms the uppermost sulphide body. Barite ore and gypsum ore ('Sekkoko') are also present and occur above, or alongside, the Oko and Kuroko deposits. Lambert and Sato (1974) point out that relative abundances of the ore types vary and in many cases one or more bed is missing.

The sedimentary rocks associated with the Kuroko deposits appear to be variable clastics ranging from mudstones to sandstones and

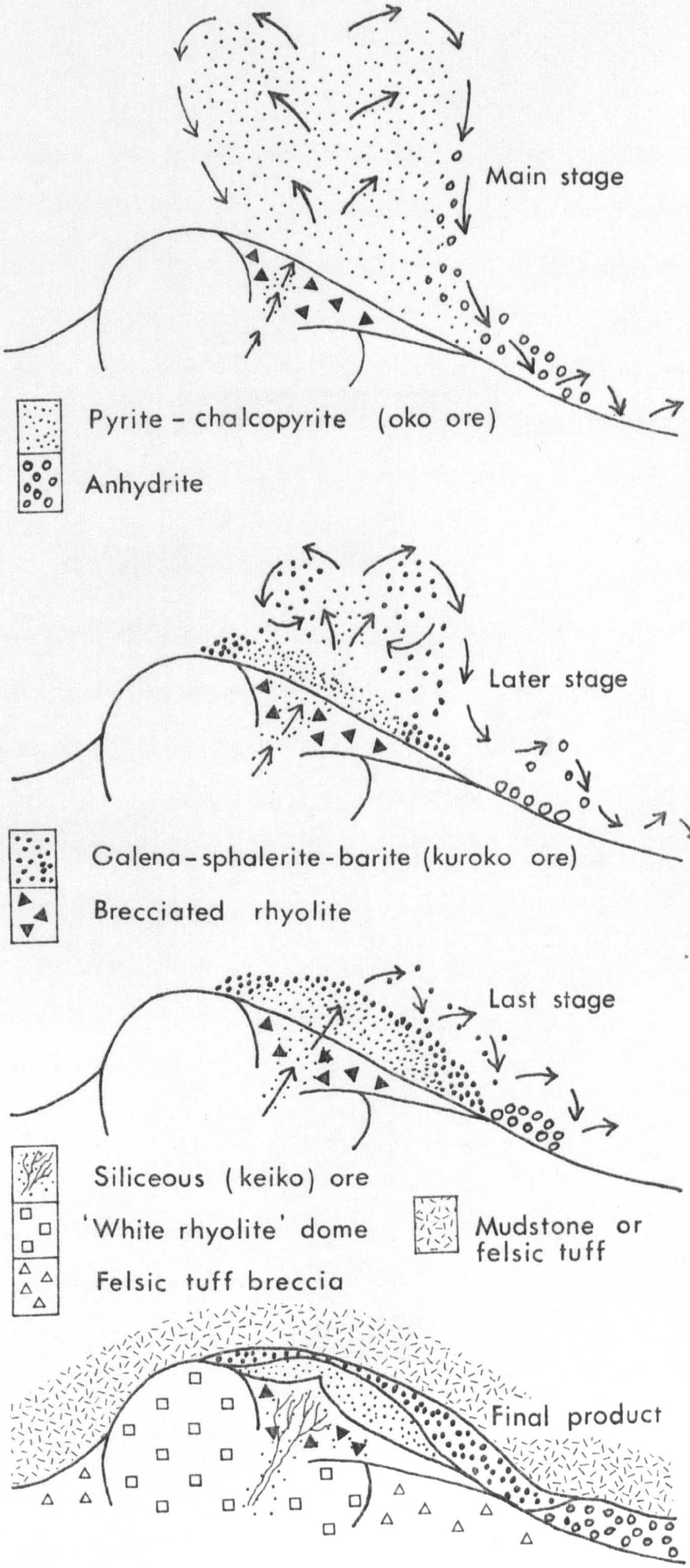


FIG. 11. 2 Schematic diagrams of the stages of Kuroko ore formation (after Sato 1972)

immature volcaniclastics, all of shallow water type. The alteration of the host rocks associated with the mineralization will be discussed in the section concerned with the Eiker deposit (11.2.4). The Kuroko deposits thus form a wide range of ore types, all of which are not necessarily developed in one area. Essential to these deposits, however, is the intimate association of felsic volcanic rocks in a stratiform relationship with a sulphide ore body which may be composed of Fe-Zn-Pb-Cu, Fe-Zn-Cu or Fe-Cu: all are submarine in origin.

Metamorphism, especially on a regional scale, might be expected to obliterate the majority of features which characterize the modern Kuroko ores. However, based on ore-mineralogy and host rock chemistry, many ancient deposits have been suggested to be of Kuroko-type. Lambert and Sato (1974) give several examples of various ages. These include the Archaean Superior deposits of Canada, the Proterozoic Skellefte deposit in Sweden, the Palaeozoic Rio Tinto deposits of Spain and Portugal, the Palaeozoic ores of New Brunswick in Canada and the U.S.A. and the Palaeozoic Australian deposits at Rosebury, Mt. Lyell, Captain Flat, Woodlawn and Mt. Morgan. Further examples are given in Tables 11.2, 3 and 4. To these may be added the deposits of the Flin Flon-Snow Lake area, Canada, as described by Koo and Mossman (1975), the Bleida deposit, Morocco, as described by Leblanc and Billaud (1978) and the Mattagami Lake deposit, Canada (Roberts, 1975).

There is a large amount of variability within these ore deposits as regards mineralogy and chemistry, particularly the presence or absence of Pb and pyrrhotite. Hutchinson (1973) considered that the lack of Pb, together with a dominance of chemically deposited sediments, was indicative of early eugeosynclinal or early subduction environments while the presence of Pb, with a dominance of epiclastic sediments, indicated late eugeosynclinal or late subduction environments, typical of Kuroko deposits. Pb however, may or may not be present for several reasons. Hutchinson himself considered that Pb was probably concentrated in the crust and that melting of lower crustal material rather than upper mantle (see Chapter Twelve) formed the Pb bearing deposits.

Lambert and Sato (1974) considered that the low Pb values possibly result from the fact that andesites have low Pb/Cu + Zn values and that the ores were formed either from fractionation of such a magma or from the leaching of andesitic hosts. Precambrian ores are often lacking in gypsum and barite deposits. Sangster (1972) considered that Precambrian seas had low sulphate contents possibly due, as Hutchinson (1973) suggested, to reducing conditions in the early Precambrian environment.

Many Precambrian ores contain pyrrhotite and, in some cases, this is considered primary (e.g. Lambert and Sato, 1974). Vokes (1962) noted that in the Norwegian Caledonides there were pyritic and pyrrhotitic stratiform sulphides and Vokes and Gale (1976) stated that most ores in low grade metamorphic areas were of pyritic type while those of higher grade areas had greater amounts of pyrhotite.

In view of the variability that has been shown to exist in Kuroko type ores in Japan and within ancient stratabound sulphides associated with acid volcanics, it could be suggested that the Grøsli deposit is of similar affinity. The question of whether the ore was deposited in a Precambrian arc environment similar to that developed in Phanerozoic plate tectonics will be discussed in Chapter Twelve.

The lack of Ni in the sulphides argues against a basic segregational origin from a plutonic mass, while the presence of high values of Zn and the absence of supracrustal basic material in the supracrustal pile is strong evidence against a deposit as formed at constructive plate margins. Fig. 11.3 shows a plot of Cu against Zn for island-arc and constructive margin ores. It can be seen that the Grøsli ore falls into the former group. The recognition of the Kuroko type of environment is dependant on ore chemistry, volcanic host rock type and sedimentary host rock type. Since high grade metamorphism in the Grøsli area has affected original supracrustals, unequivocal arguments cannot be made for the syn-volcanic origin (see remarks on the hydrothermal replacive alternative).

11.3.4 <u>Possible environment of the Eiker deposit</u>

The origin and depositional environment of the ore deposit at Eiker is more speculative owing to the nature of the enclosing supracrustals. Once again, the Ni/Cu ratios of the ores are too low and not consistent for a basic magmatic origin. However, depending on the precise interpretation of the interbanded supracrustals several environments of deposition can be envisaged within a general volcanogenic regime. Three questions arise concerning the origins of the supracrustals. Firstly, whether the amphibolites in the southern sub-area were laid down with the proto-gneisses or represent later intrusions. Secondly, whether the gneisses themselves are entirely volcanic (or volcaniclastic) or part sedimentary (i.e. continentally derived material). Thirdly, whether the Mg-rich rocks represent metamorphosed sediments or hydrothermally altered sedimentary and/or igneous material.

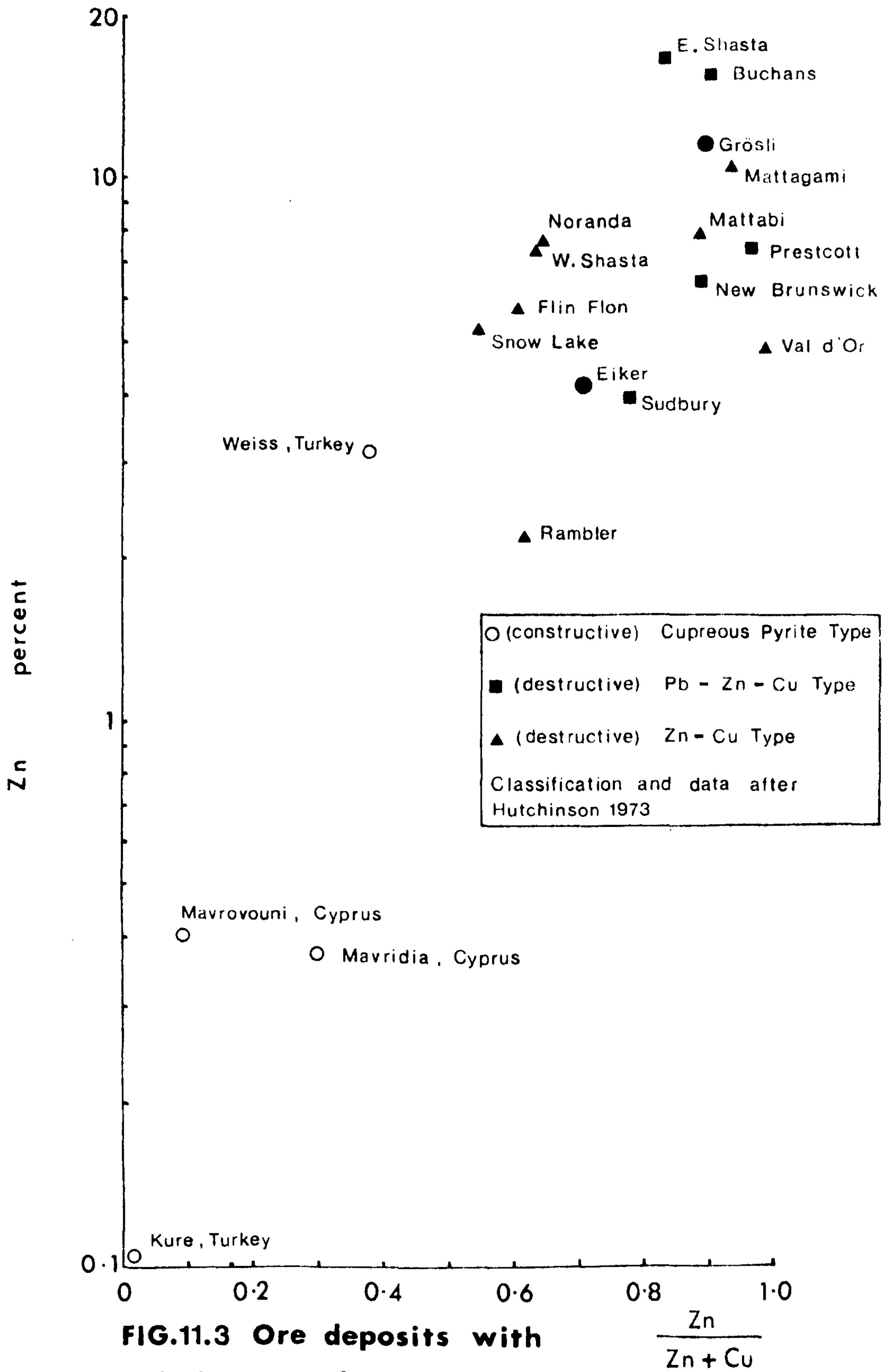


FIG.11.3 Ore deposits with affinities to destructive or constructive plate margins

a) The Amphibolites

Field and major element data suggests that the southern sub-area amphibolites are part of the supracrustal sequence, whereas the trace element data suggests that they may be later intrusives (see section 9.5).

Strike and dip gradations from gneiss, through hornblende gneiss, to amphibolite (in the southern sub-area) are taken as representing original facies changes in the supracrustal series: hence the amphibolites are seen as an integral part of the supracrustal sequence. The amphibolites of the Eiker area can be seen from Table 9.1 to show variations from basic to more intermediate compositions.

The interpretation of the origins and environments of the emplacement of these amphibolites is difficult, especially since some conflicting evidence is present in the comparison of trace and major element data. The relative ages of the northern and southern sub-area groups cannot be unequivocally demonstrated since they are nowhere in contact with each other. If the southern sub-area amphibolites are considered to be part of the supracrustal pile then they may be seen as the more intermediate members of a fractionation series, which might also have produced the Grøsli acidic gneisses. Support for this idea is evident in Fig. 11.4. One problem, however, about such an affinity is the fact that the possible Grøsli-Eiker trend line cuts the tholeiite/calc-alkaline boundary as defined by Irvine and Baragar (1971). Mobility of alkalies during metamorphism may have played a part in producing the present trend. If this trend is valid, the most obvious factor is one of iron depletion (which becomes stronger with fractionation) and no iron enrichment. This trend has been used by Miyashiro (1974)

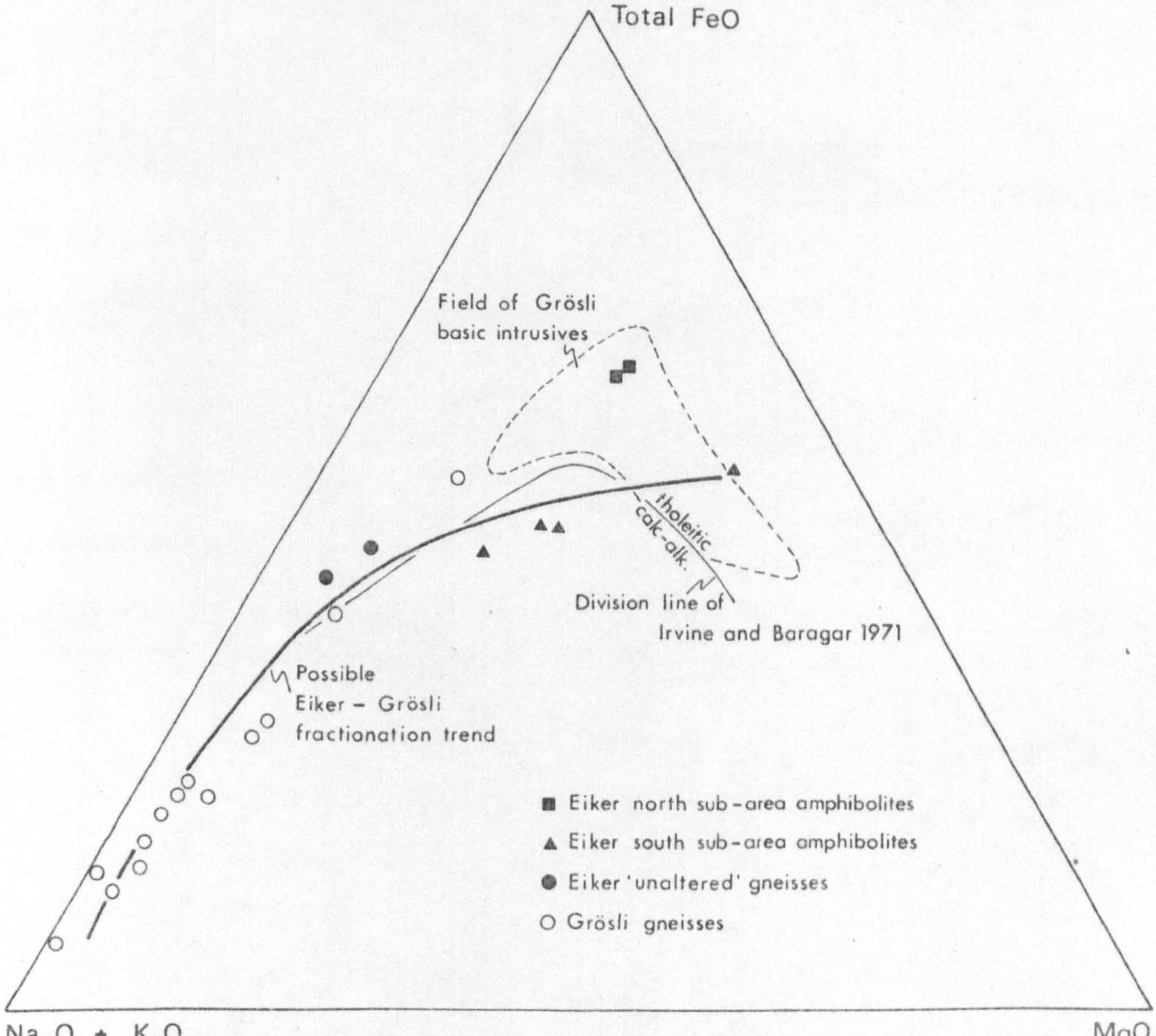


FIG. 11.4 Relationships between rock groups from Grösli and Eiker.

to characterise calc-alkaline series of volcanic island arc rocks. The northern sub-area amphibolites on this plot occur well within the tholeiitic sector of Irvine and Baragar (1971) and appear to have greater links with the younger Grøsli 'Vinor' bodies. Thus, in such a case, the southern sub-area amphibolites would be older than those of the northern sub-area and of a different chemical lineage.

Jacobsen and Heier (1978) dated the metamorphism of the quartzo-feldspathic gneisses and interbedded amphibolites in the Kongsberg series at 1600-1500 m.y.B.P., whilst the intrusion of dioritic gneisses was dated at about 1520 m.y.B.P.: these were considered by Jacobsen (1975) to be of calc-alkaline affinity. He considered that the sequence of events in the Kongsberg series was as follows:

a) $\approx$ 1600. Deposition of island-arc sodic volcanics (basalts rhyolites) and some sedimentary rocks.

b) $\approx$ 1600 to 1500. Intrusion of calc-alkaline gabbroic to dioritic rocks (1520): deformation and metamorphism (Sveco-fennian).

c) $\approx$ 1370 to 1200. Intrusion of the 'Vinor' gabbros (olivine tholeiites).

d) $\approx$ 1200 to 1100. Second metamorphic episode (Sveconorwegian).

The above sequence is in agreement with a $\approx$ 1600 m.y.B.P. age for the southern sub-area amphibolites and a $\approx$ 1370 to 1200 m.y.B.P. age for the northern sub-area intrusives.

Information conflicting with that of the AFM diagram (Fig. 11.4) is presented in the form of trace element plots of the amphibolites (Chapter Nine). The southern sub-area rocks appear to have 'mid-ocean ridge basalt' (MORB) characteristics, but with trends towards 'within plate basalt' (WPB) types. The northern sub-area amphibolites appear to be transitional 'island-arc basalt'/MORB in their characteristics. Thus the AFM diagram and trace element data suggest differing origins for the two sub-area amphibolites. These two sets of data and field relationships strongly suggest that the northern and southern sub-area amphibolites are of different age and/or origin.

Considering the trace element data, the southern sub-area amphibolites are of MORB/WPB type which have been identified by Pearce and Gale (1977) and Pearce (in press) as the products of rifting on continental margins similar to the Red Sea environment at present. The two northern sub-area amphibolite analyses appear to be of IAB composition but with transitional natures towards MORB characteristics described in rocks produced at divergent margins above subduction zones (Pearce and Gale, 1977). If the two groups are seen as being genetically linked, then they can be related to the incipient opening of a back-arc basin above a subducting plate. Initially magmatism could have occurred, cutting continental material (the metamorphosed supra-crystals at Eiker) and producing the southern sub-area amphibolites. Further rifting may then have been associated with the northern sub-area proto-amphibolite intrusions which have characteristics more in common with Cyprus type (IAB/MORB) volcanics. The amphibolites at Eiker, in this case, would be genetically unrelated to the ore deposit and older then the Grøsli basic intrusives. There are problems associated with this suggestion in that there is no evidence to support the existence of an island arc environment after the Svecofennian. The southern sub-area amphibolites have trace-element characteristics similar to those of the Grøsli 'Vinor' intrusives (as shown in a comparison of Figs. 4.10 and 9.7). This is at variance with the AFM diagram (Fig. 11.4) which, although based on a small number of analyses, suggests a calc-alkaline affinity for the southern-sub-area amphibolites. With the northern sub-area amphibolites the trace element and the AFM data are equally at variance. The AFM data suggests a similar affinity to the 'Vinors' of the Grøsli area while trace element data suggests links with an island-arc environment.

It is concluded that, in the absence of greater numbers of analyses for the Eiker amphibolites, the chemical data must be viewed with caution, and that no definite conclusions can be drawn from the analyses alone. Further, plate tectonic mechanisms of the present may have had very different counterparts in the Precambrian and trace element data may be indicative of somewhat different environments. The discordant contacts and intrusive shapes of the northern sub-area amphibolites obviously indicate these are later injections in the supracrustal gneisses. The question of whether these are 'Vinor' intrusives or earlier rocks cannot be answered in the absence of radiometric dating. The amphibolite body to the north of the shear zone at Eiker has been involved in high grade shearing which as responsible for the cataclastic textures in the supracrustals. Starmer (1977) stated that the main cataclasis at the Kongsberg Series western margin was almost complete when earliest 'Vinor' bodies were injected. Thus, if the high grade cataclasis is of the same age and if the northern sub-area amphibolites are 'Vinor' intrusives then they must represent very early intrusions of this phase. Such an age is supported by the position of the two analyses on the AFM plot.

b) The Gneisses

Most of the intermediate and acidic gneisses plot well away from igneous trends (Chapter Nine). The obvious conclusion is these rocks were sediments such as Mg-rich shales, together with some greywackes or sandstones. Their relatively low Ca contents preclude dolomitic precursors as a source of high Mg values. However, it is possible that these rocks represent hydrothermally altered metamorphic equivalents of volcanic material. Indeed the Zr ppm versus Niggli si plot (Fig. 9.3) shows a very pronounced negative trend which suggests that these rocks represent primary igneous rocks or volcaniclastic deposits.

If these rocks are compared with the two analyses (9 and 10) which do appear to be of volcanic affinity, then the main chemical differences are an increase in Mg, with a decrease in Ca and Na. These rocks are variable quartz-sericite-sillimanite-biotite-cordierite-garnet gneisses.

Latvalahati (1979) working on Cn-Zn-Pb stratabound ores from the Aijala-Orijarvi area, S.W. Finland, stated that an areal "blanket" type of alteration existed around the ore deposits, consisting of cordierite bearing sericite and muscovite schists and gneisses, which were considered to be the more acid volcanics of a Kuroko type ore deposit. Chemically the altered rocks have increased Mg, Fe, Mn and Ti and are depleted in Na. Huhtala (1979) and Helovuori (1979), studying similar ore deposits from the Pyhäsalmi and Pielavesi district in Finland also described a sequence of altered host rocks. Here, as a result of Mg addition, the acid volcanics and tuffaceous sediments had recrystallized as cordierite and garnet-cordierite gneisses. Cordierite-anthophyllite-garnet gneisses, anthophyllite-amphibolites and cummingtonite gneisses are the result of metamorphism of altered mafic tuffs. Chemically, they stated, Ca and Na have been depleted while Fe, Mg and Al have increased in the alteration. Berge (1978) concluded that Mg concentration, in association with massive sulphide deposits of submarine volcanogenic origin, may be due to the reaction of volcanic rock with either sea water, or hydrothermal solutions accompanying volcanism, or both.

Koo and Mossman (1975) showed that the wall rocks in the Flin Flon mine were Si, K and Na depleted, with associated Mg and Ca enrichment. They believed that the alteration resulted partly from metasomatism induced by shearing and partly from hydrothermal ore solutions. Alteration of wall rocks at Mattagami Lake (Ontario) has resulted in the addition of Mg and Fe and a decrease in alkalies and sometimes Si.

Lambert and Sato (1974), Tatsumi et al (1972) and Sangster (1974) all give examples of alteration associated with Kuroko-type deposits. Four zones are recognised; an outermost montmorillonite-zeolite zone is followed by a sericite-chlorite-pyrite zone and a sericite-chlorite-quartz zone. Finally, closest to the ore body, there is a silicified zone containing some sericite and chlorite. Mg is enriched and Na is depleted in all zones: Ca is depleted in the zones adjacent to the ore body while Fe is slightly enriched in such zones.

Stanton (1976) stated that the sillimanite gneisses of Broken Hill, Australia, could be accounted for by the addition, to a dacitic parent, of kaolinite and silica (associated with ore solutions): Ca and Na impoverishment accompanied by Fe, Mg and K enrichment could be due to clay mineral precipitation and feldspar breakdown again associated with hydrothermal activity. In this respect, the resultant rocks do not just represent hydrothermally altered volcanics, but hybrids formed by the mixing of ore solution silicates and volcanic rocks altered by the same solutions.

The rocks at Eiker, if they are considered to be originally volcanic, show element enrichment and depletion patterns which are totally consistent with the Kuroko type alteration patterns. This alteration is not developed at a distance from the ore deposit in the Eiker area. The single gneiss analysis (no. 2) which plots on the amphibolite trend in the Zr ppm versus Niggli si plot (Fig. 9.3) is a garnet-hornblende-biotite gneiss and could possibly represent one of the bedded amphibolites which has been hydrothermally altered.

The Mg-rich supracrustals are thus considered essentially volcanic rocks which have been hydrothermally altered. The rocks are essentially the same as those in the Grøsli region but are less acidic, representing rhyo-dacitic and dacitic compositions rather than rhyolitic rocks.

11.4 A HYDROTHERMAL ORIGIN

It has been shown in the last section that the ores may be considered as a combination of sedimentary, hydrothermal and volcanogene in the sense that they are submarine deposits laid down from hot solutions of volcanic origin.

Hydrothermal fluids can be defined as "heated aqueous fluids without implication as to origin" (Barnes, 1967). The term has already been used thus in the last section. However, in this section, it is utilised in a more strict sense to describe post-depositional host rock replacement with no volcanic associations. There are several periods in the history of the Grøsli and Eiker areas when the introduction of the ores could have caused replacements of host rocks. (The only definite constraint on the date of the ore deposition is a pre-'Vinor' age with the subsequent high grade metamorphism.)

Firstly, the ores may represent pre-Svecofennian replacements. In the absence of any categorical evidence against such an early hydrothermal replacive nature, it is important to emphasise the numerous features and similarities with Kuroko-type deposits (i.e. suggesting the ores were laid down with the supracrustals). The same argument can be applied to any suggestion of a post-Svecofennian, pre-Sveconorwegian replacement.

At Eiker, the shear zone is considered to be of the same age as the mylonite zones (at the Kongsberg Series western margin) and therefore the ore body here was formed before the Sveconorwegian orogeny. It could be argued that the shearing event facilitated the introduction of the ore fluids. There is little evidence against such a process, except the alteration features of the acidic supracrustals (which conform well with Kuroko-type alterations). It is suggested that in fact the ore was present before this shearing and represented a structural discontinuity which faciliated fracturing and shearing.

Extensive metamorphic mobilization of pre-existing ore bodies is considered unlikely, although concentration of disseminated material by metamorphically activated fluids is considered a strong possibility by some workers (e.g. Mookherjee, 1970; White, 1968 and Sangster, 1971). In the case of Grøsli and Eiker, microscopic evidence shows textures which are the result of metamorphic modification rather than mobilization of an existing ore body. Thus, in conclusion, hydrothermal replacement must be considered as unlikely and the weight of evidence suggests a volcanogenic origin.

11.5 A SEDIMENTARY ORIGIN

Considering the fact that the supracrustals were largely acid volcanics and without the characteristic lithologies associated with sedimentary sulphide deposits such as the Roan and Mansfeld occurrences (as classified by Stanton, 1972 - see Table 11.2), the Grøsli and Eiker deposits do not appear to result from sedimentation without associated volcanism.

CHAPTER TWELVE

ENVIRONMENTAL MODELS OF ORE FORMATION AND CONCLUSIONS

12.1 INTRODUCTION

From the evidence presented and discussed in Chapter Eleven, it is concluded that the deposits at Grøsli and Eiker most likely represent volcanogenic ores formed in Kuroko-type environments. The two deposits are now less than 30 km apart and despite folding, sub-vertical litho-banding and cataclasis in some places, they could have formed in the same sub-marine basin and be related to the same volcanic suite, with increasing fractionation (acidity) of the rocks from Eiker to Grøsli.

This chapter reviews the comparability of modern continental margin regimes with those thought to be operating in the Proterozoic, with particular reference to ore depositional environments. The characteristics of proximal and distal ore deposits of Kuroko-type and the effects of metamorphism on these characteristics will also be reviewed and discussed in terms of Grøsli and Eiker.

12.2 PROTEROZOIC PLATE TECTONICS

The subject of crustal evolution in the Proterozoic is too large for discussion in the present context. The principal point of interest is whether ore deposition occurred in island-arc environments where active plate subduction was taking place. There are two opposing schools of thought on Proterozoic 'plate tectonics'. The first maintains that plate movement took place in much the same way as it does at present. The second proposes that little or no horizontal movement of plates took place throughout the majority of the Proterozoic.

Windley (1973) noted that Precambrian sediments, igneous intrusions, metamorphism, structures and mobile belts were significantly different from those of the Phanerozoic. He stated that since this is so, it might be expected that types of plates and plate movements were also different. He concluded that there is little evidence for extensive ocean-floor spreading in the Proterozoic and favoured intra-continental 'plate' movements as the only manifestations of tectonics. Ringwood (1970) suggested that two processes have operated throughout geological time, to generate new crust. Firstly, subduction and fractional crystallization of new oceanic crust and secondly, vertical mass transport involving fractional melting of mantle material and the rise of magmas into the crust. Ringwood considered the second method to be more common in the Precambrian. Windley (1973) concluded that many of the mobile belts of the Proterozoic represented rifting of narrow intra-continental belts associated with the magmatism of Ringwood's vertical differentiation process. In contrast with the above opinions, Hoffman (1973), after studying the Coronation geosyncline in the Canadian shield, considered that the similarities with Phanerozoic Cordilleran-type geosynclines were so strong that modern plate-tectonics must have been operative in early Proterozoic times. Van Schmus (1976) also considered the processes similar to modern plate tectonics may have occurred early in the Proterozoic.

Glikson (1976) considered that no ocean domains were developed during the early and mid-Proterozoic (2500 to 1000 m.y.B.P.) and that consequently no continental drift occurred during this time span. He stated it was possible that a world-wide sialic crust existed during this period, formed by the nucleation of proto-continents.developed by recycling of the earlier (Archaean) oceanic crust. This cycling may have been in the form of down-buckling, or subduction, with resultant partial melting to form sialic material.

At the end of the Archaean (2600 m.y.B.P.) a supercontinent containing Africa, South America, North America and the Fennoscandian shield (Piper, 1976a) formed from the nucleation of smaller proto-continents. Glikson (1976) suggested that this nucleation occurred as a result of cratonization, in situ, rather than as a result of any extensive drifting and continental collision while Piper (1976b), on the basis of palaeomagnetic evidence, maintained that this super-continent then remained intact and static until it began to break-up around 1200 - 1000 m.y.B.P. (in Grenville-Sveconorwegian times). Metamorphism (including the Svecofennian) would thus have been related to high heat flows occurring within ensialic mobile belts.

Addition of crustal material in such an environment would be largely by vertical differentiation, possibly through rifting systems as suggested by Glikson (Fig. 12.1) or by upwelling silicic currents as suggested by Shackleton (1973). This process has definite analogies with the origins suggested by Hutchinson (1973) for Archaean-Proterozoic stratabound ore deposits. He provided two models for the generation of Zn-Cu deposits in the Archaean and Pb-Zn-Cu deposits in the Proterozoic. He considered that the Zn-Cu ores were derived from degassing of the mantle while their associated volcanics were the products of a poorly differentiated mantle of intermediate average composition

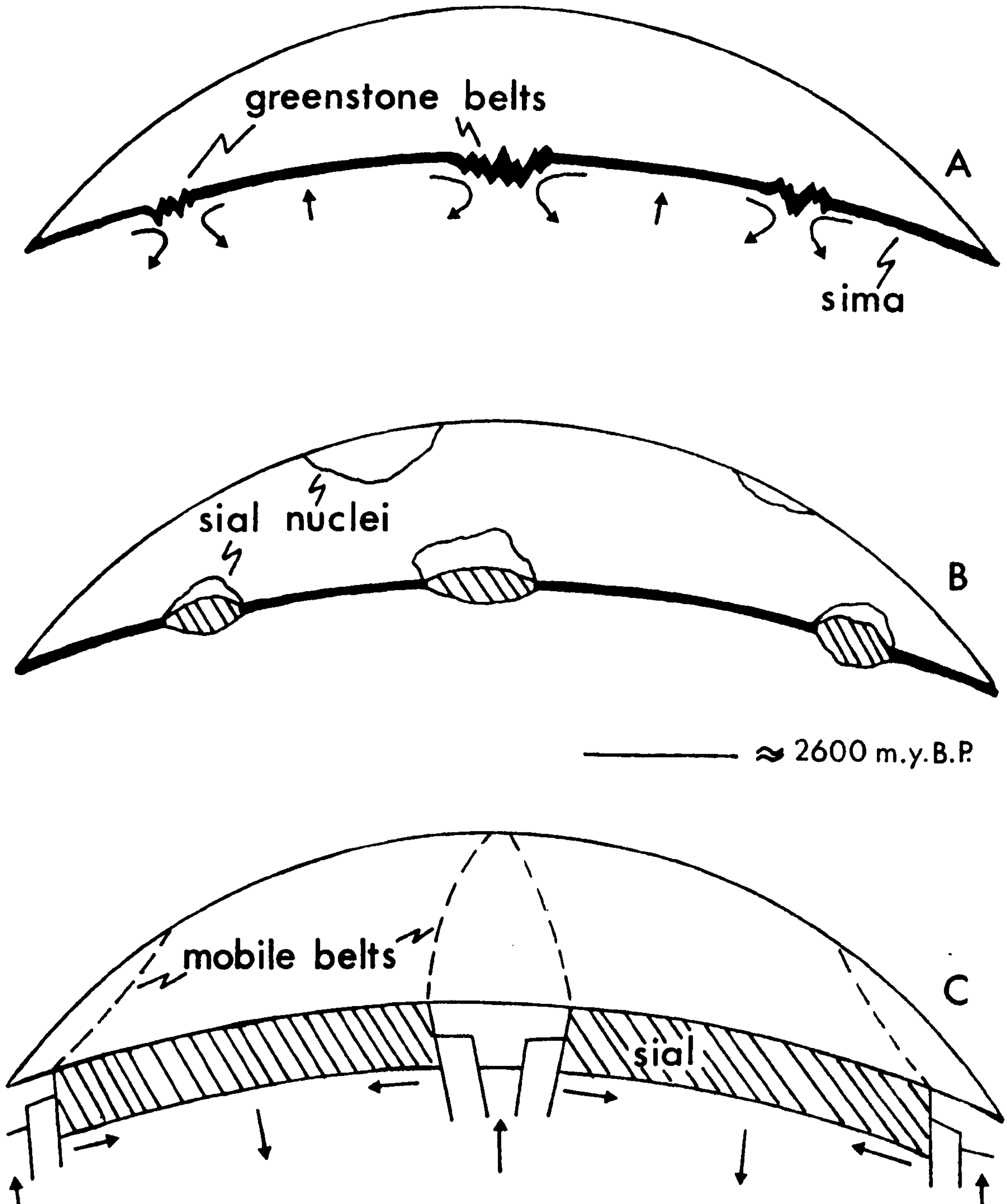


FIG. 12·1 Crustal development in the Archaean and Proterozoic. (after Glikson, 1976)

Development of downbuckling in the Archaean (A) gave rise to the formation of proto-continents (B) which then coalesced in situ to form major Proterozoic continents (C). Subsequent ensialic activity produced mobile belts.

(Fig. 12.2). The Pb-Zn-Cu ores were associated with volcanism, derived from the anatectic melting of a sialic crust which was much thicker in the Proterozoic (Fig. 12.3).

The question of whether the Kongsberg Series magmas developed from thick Proterozoic crust, or directly from differentiation of the mantle is considered in section 12.4. The important factor, however, is that the vertical mass transport process (without plate-tectonics and subduction) appears to be a possible mechanism for the origin of the Grøsli and Eiker deposits.

Much research has suggested that continental drift did occur during the Proterozoic. Much argument has revolved around the Grenville mobile belt and it has been suggested that it was the result of continental collision (Dewey and Burke, 1973). Palaeomagnetic evidence from Irving and McGlynn (1976) suggested that continental collision occurred, while Baer (1976) and Wynne-Edwards (1976) both suggested that rifting was followed by contraction and collision to form the Grenville belt.

Chase and Gilmer (1973) likewise considered that the mid-continent gravity high running from Lake Superior to Kansas was associated with volcanics similar to ocean tholeiites and gabbros. They considered that the rifting was due to the oblique collision of the Grenville plate with the North American plate 1300 - 1100 m.y.B.P. as suggested by Donaldson and Irving (1972). Hietanen (1975) also suggested an island-arc environment for the Fennoscandian shield in the Proterozoic (Fig. 12.4). Sutton (1976) considered that both ensialic orogenesis and sea floor spreading (with destruction of oceanic crust) occurred during the Proterozoic.

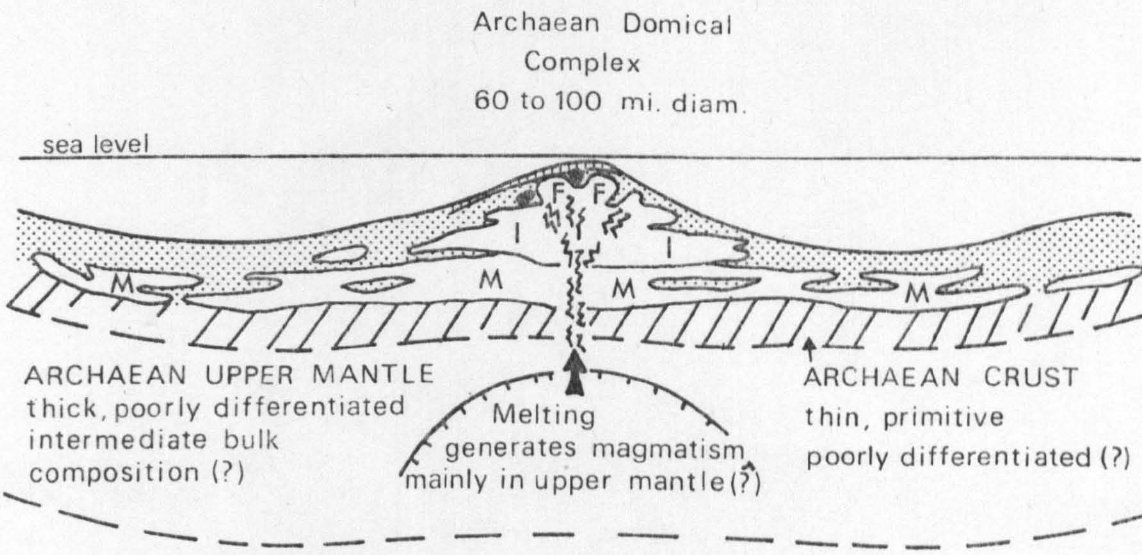


FIG. 12·2 Zn – Cu type of volcanogenic massive sulphide deposit in Archaean tectonic setting. (after Hutchinson, 1973)

F Felsic
I Intermediate } volcanics
M Mafic

Iron Formation – sulphide facies

Zn – Cu body

Volcaniclastic and chemical sediments

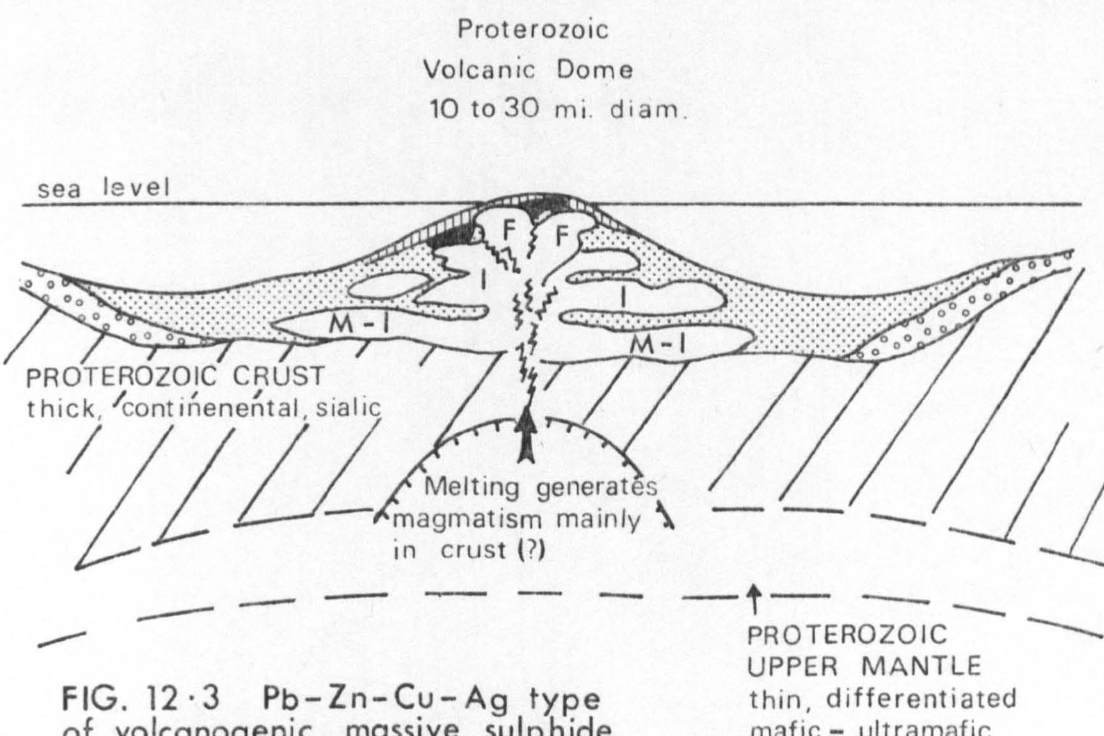
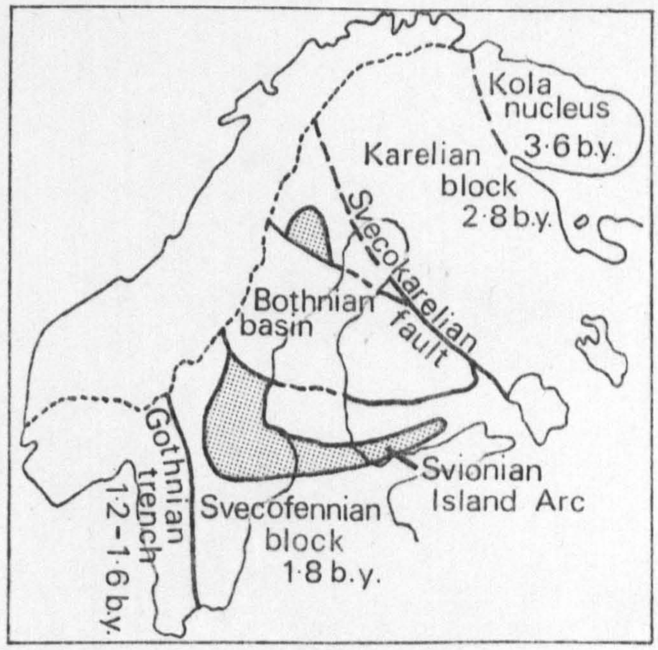


FIG. 12·3 Pb–Zn–Cu–Ag type of volcanogenic massive sulphide deposit in Proterozoic tectonic setting. (after Hutchinson, 1973)

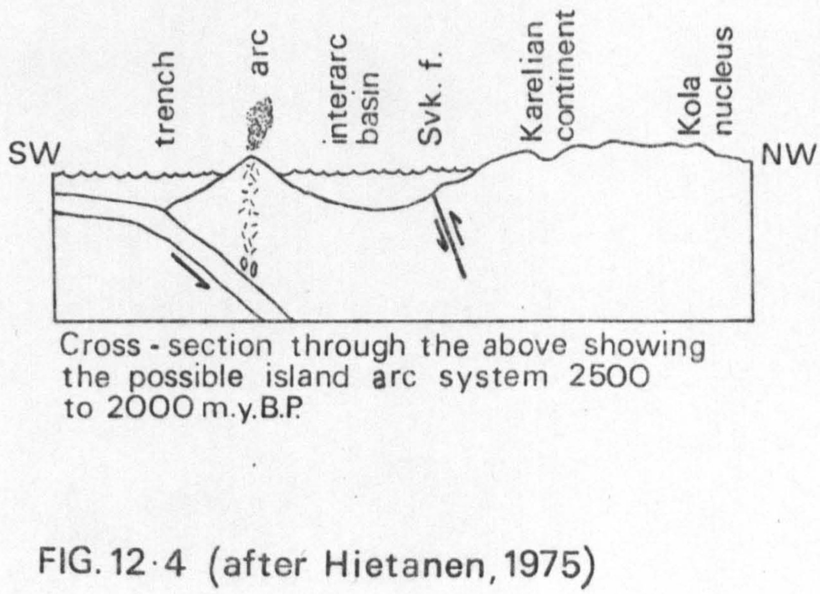
Pb – Zn – Cu – Ag body

Fine epiclastic, volcaniclastic and chemical sediments

Coarse clastics



Geotectonic elements and their radiometric ages in the Fennoscandian shield



Cross - section through the above showing the possible island arc system 2500 to 2000 m.y.B.P.

FIG. 12·4 (after Hietanen, 1975)

The above resume indicates that there is still active argument over the role of plate tectonics in the Proterozoic. Windley (1977) noted that it seems possible that early Proterozoic Cu-Pb-Zn deposits of volcanogenic origin formed in the first active continental margins similar to those of the Mesozoic. However, he concluded that "the role of plate-tectonics during the period 2500 - 1000 m.y. ago is poorly understood and thus controversial".

It is generally agreed that the Proterozoic supercontinent started to break-up 1200 - 1000 m.y. ago and that this separation continued until around 850 m.y. ago. In the Grenville province this break-up was accompanied by the intrusion of many igneous rock types including basalts, dolerite dykes, Rapikivi granites, alkaline complexes and carbonatites.

The 'Vinor' basic intrusives may represent equivalents in the Fennoscandian block. That the separation was short lived in this area is evinced by the fact that the Sveconorwegian orogeny took place very shortly afterwards. The 'Vinor' intrusives (and associated 'hyperites' of the Bamble region) may well have been associated with the continental rifting of Grenvillian age (as described by Baer, Wynne-dwards, Chase and Gilmer). The rifting failed fairly soon after its conception and the continental crust was involved in the Sveconorwegian orogeny after relatively little drift.

In conclusion, there is good evidence both for and against 'modern' plate tectonic movement in the Proterozoic. Consequently, origins for stratabound Cu-Zn-Fe sulphides of volcanogenic origin can be thought of both in terms of Kuroko-type and vertical differentiation type models.

12.3 PROXIMAL AND DISTAL CHARACTERISTICS OF KUROKO DEPOSITS

The spatial relationship of the ore deposit to the volcanic vent is considered in this section. The characteristics of proximal and distal deposits are taken from work on Kuroko-type areas. However, the features might be considered to develop in any submarine environment where acidic volcanogenic activity is occurring. Thus, characteristics may be expected to be similar in both the genetic models outlined in the last section.

The Grøsli and Eiker deposits are now less than 30 km apart. Their original spatial relationship is extremely difficult to assess. The extent of crustal shortening in the Kongsberg Series has not been estimated, but may well be considerable, in view of the fact that the series has undergone two major orogenic episodes. The respective stratigraphic level of each deposit is likewise unknown, and if transposition of the litho-banding has taken place, the deposits might even be lateral equivalents. However, since silicate-rock lithologies and the relative ages of the ores show some similarities in both areas, the two deposits may be genetically linked and may even have been deposited in the same marine basin. In this case, determination of proximal and distal characteristics would indicate the position of the volcanic centre associated with the ore forming solutions, assuming that the two ore deposits are not transposed lateral equivalents.

Colley (1976) examined undeformed Kuroko stratabound deposits from Fiji, and identified five types of deposit characterised by their proximity to the exhalative centre. The spatial relationships of each type are shown in Fig. 12.5. Type I deposits are deep level feeder systems in the form of stockwork veins. Type II deposits occur where

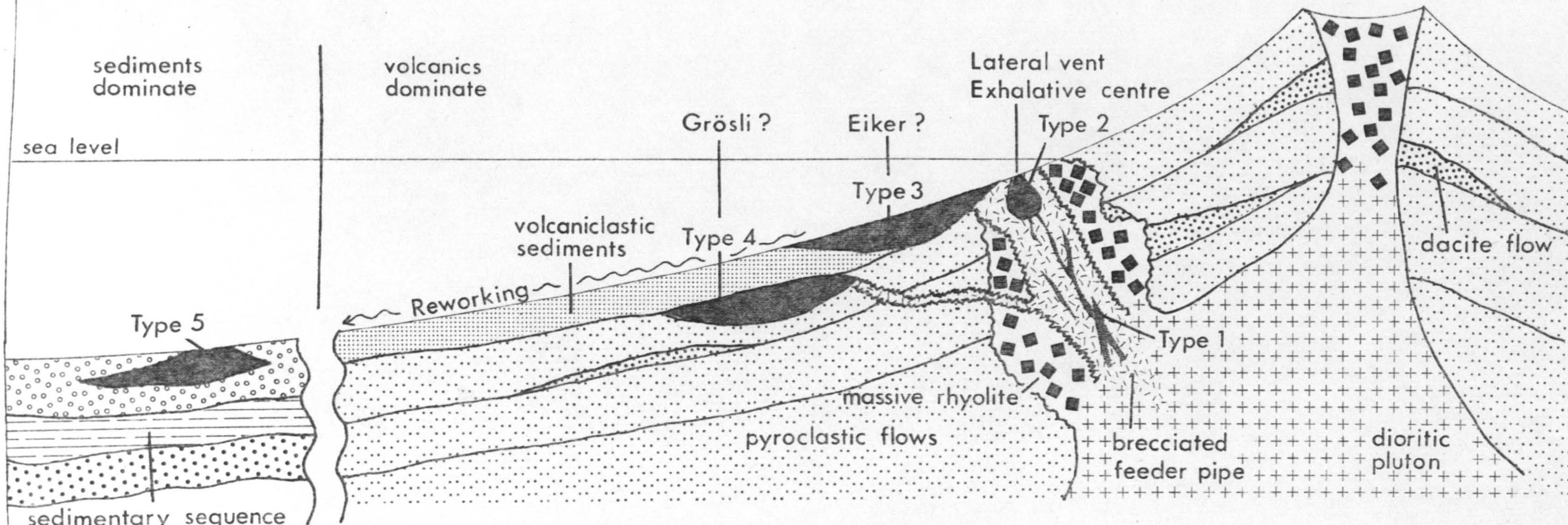


FIG. 12·5 Model showing position of various kinds of Kuroko-type deposits within a volcano-sedimentary succession. (after Colley, 1976) the feeder complex reaches the surface and forms an exhalative centre. The ore bodies are in the form of small pods or pipes which pass vertically down into stockwork ore. They occur in the immediate vicinity of the feeder vent. Type III deposits (including the 'classic' Kuroko bodies) have an almost identical setting to the type II deposits except that the ore body is near horizontal and lenticular. Colley considered these deposits to be formed on the slopes surrounding the exhalative centre. Stockwork ore and altered country rock around feeder channels occur below the ore body.

Type IV deposits are also stratabound but occur away from feeder complexes and result from ore solutions moving away from the eruptive centre before precipitating their metals. Colley also pointed out that these deposits could form from weak and diffuse fumarolic activity so that no definite alteration channels were formed. Type V deposits are similar to type IV ores in being remote from a feeder complex. They may form in the same way as type IV deposits, but at even greater distances from the exhalative centre, or they may form from the reworking of other deposits (types I to IV) which were then deposited in deeper marine basins. In either case, type V deposits may be expected to occur associated with predominantly sedimentary rocks, while the first four types occur in an essentially volcanic host rock.

In this classification, types I, II and III deposits can be considered proximal while type V deposits are obviously distal. Type IV deposits could fall into either category.

Large (1977) discussed proximal and distal ore deposits in terms of their chemical evolution. He considered that temperature, pH, fO_2 and concentration of total dissolved sulphur in the ore fluids were the important factors controlling the behaviour of that fluid and its ability to hold base metals. He stated that there were two distinct environments of ore transport. Firstly, reduced, high temperature (above 275^{o}C) conditions in which the fluid pH was acid to neutral. Secondly, oxidising conditions at any temperature and any pH. Proximal ores were deposited when ore solutions mixed with sea water, which resulted in a change from the first to the second environment. Distal ores were deposited from lower temperature, oxidising solutions within the second environment which suffered a drop in fO_2 or increase in temperature.

As a result of experiments on mineral-solution equilibria in the Fe-S-O system, Large (1977) concluded that banded pyritic Pb-Zn-Cu ores ranged across the proximal-distal division and generally resulted from direct precipitation onto the sea floor. He also compared the Archaean Cu-Zn and Kuroko Pb-Zn-Cu deposits and considered that their ore solution chemistries were probably very similar except for two points. Firstly, the absence of Pb in Archaean deposits which was due to the lack of this element in the ore forming fluids. (Zn and Pb behave very similarly under varying physico-chemical conditions and any Pb present may be expected to have formed galena along with the sphalerite.) Secondly, the predominance of pyrite in Kuroko type ores and of pyrrhotite in Archaean ores, was probably a result of lower total dissolved sulphur contents in the Archaean solutions.

Large (1977) concluded that Archaean ores were all of proximal type, since the seas of that period were of a reducing nature which prevented the transport of base metals away from the hydrothermal outlet. However, Windley (1977) considered that by 1700 m.y.B.P.

the oxygen level of the oceans was increasing rapidly due to the activity of primitive organisms such as blue-green algae. Thus Large's explanation for the absence of distal ore deposits may not have been applicable for the Proterozoic.

Windley also stated that Pb-Zn-Cu ores of Kuroko type chemistry were present by the early Proterozoic. Thus, it was probable that both proximal and distal ores were capable of being formed at this time. Accordingly, the Grøsli and Eiker deposits must be considered from both possible viewpoints.

Plimer (1978) related environment of deposition to the proximity of the volcanic centre. He stated that, with increasing distance from the vent, the Fe, Cu and S content of the base metal deposits decreased while the Zn, Pb, Ag, Mn, Ba and F content increased. The trend from Cu domination to Pb-Zn domination, he considered to be the result of increased mixing of sea water with ore fluids, away from the vent. He stated that the host rocks of proximal deposits were altered so that their MgO, FeO, S and base metal values were substantially increased, with minor additions of K_2O and SiO_2. Distal rocks, however, showed slight additions of K_2O, MnO, TiO_2, FeO and base metals, with slight depletions of MgO, Na_2O and CaO in both environments. Summaries of the change in ore character with distance from the exhalative centre are reproduced in Figs. 12.6 and 12.7. The chemistry of the ore deposits is presented in Table 12.1.

Finlow-Bates et al (1977) and Plimer and Finlow-Bates (1978) considered the dominance of pyrite or pyrrhotite in terms of depositional history. Previous authors (including McDonald, 1967; Vokes, 1969 and Lambert, 1973) suggested that pyrrhotite can be produced from pyrite

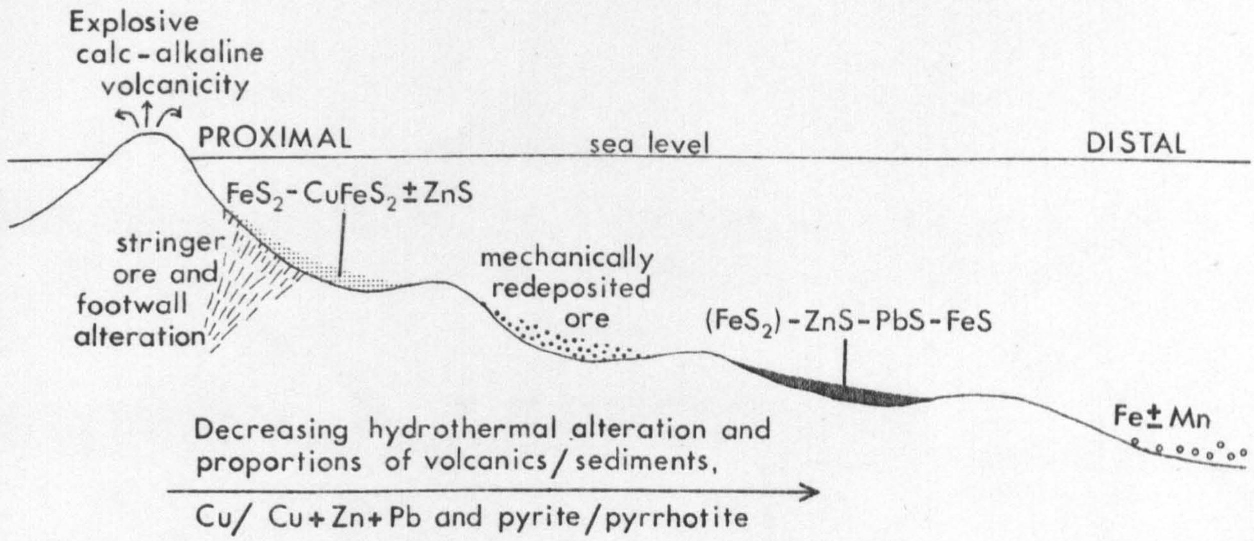


FIG. 12·6 Relationship between ore type and distance from volcanic vent (after Plimer, 1978)

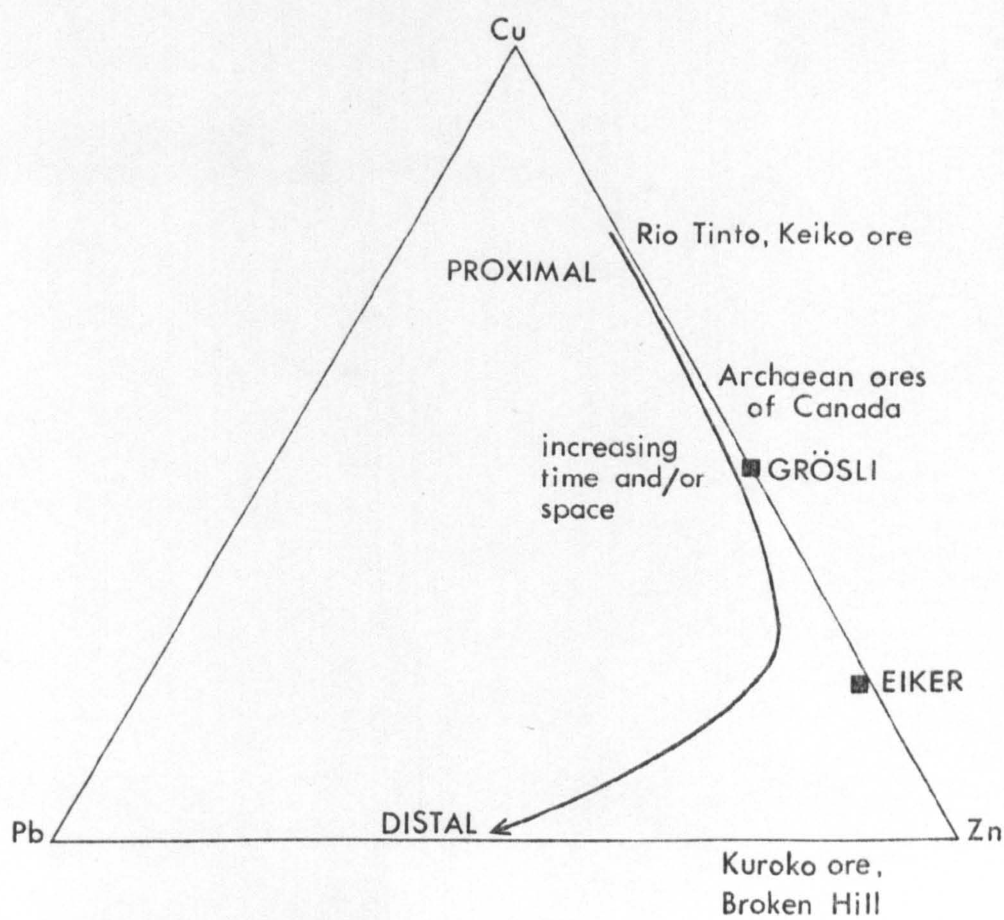


FIG. 12·7 Metal ratios in proximal and distal ore deposits (after Plimer, 1978)

TABLE 12.1

MAJOR AND TRACE ELEMENT CHEMISTRY OF THE GRØSLI AND EIKER MASSIVE ORES

(Major elements - Fe, Zn, Cu - in wt%; trace elements in ppm)

	Fe	Zn	Cu	Cr	V	Ti	Co	Mn	Ni	Cd	Pb
1	42.84	2.64	0.98	76	11	148	181	2659	26	12	N.D.
2	46.61	4.02	0.06	79	25	199	43	1333	45	6	N.D.
3	40.60	2.91	3.08	55	13	113	285	316	31	N.D.	N.D.
4	35.01	5.95	1.07	66	9	237	197	2650	28	58	N.D.
5	39.47	6.13	0.19	67	20	157	200	2575	52	31	N.D.
6	40.58	1.97	5.86	70	10	125	262	2138	19	6	N.D.
7	32.06	5.16	0.55	93	50	478	201	4211	32	N.D.	N.D.
8	26.32	0.84	15.69	88	35	546	28	1248	16	N.D.	N.D.
9	46.64	14.10	3.02	72	5	1	339	408	273	514	10
10	39.60	2.79	18.75	116	5	3	434	161	135	111	N.D.
11	44.55	9.74	1.83	56	5	N.D.	500	341	30	394	65
12	39.03	19.63	0.88	100	9	N.D.	62	228	106	640	N.D.
13	47.27	8.98	1.54	48	20	63	425	543	150	352	N.D.
14	47.92	11.13	0.59	131	25	277	69	225	314	367	N.D.
15	50.90	8.47	1.23	48	3	53	81	148	362	377	N.D.
16	45.49	11.74	0.89	93	7	N.D.	554	382	298	961	N.D.
17	42.57	9.96	1.55	374	88	1262	56	891	231	333	N.D.
18	39.98	16.73	0.36	111	7	N.D.	66	296	261	502	N.D.
19	53.54	3.12	0.47	58	4	N.D.	59	340	146	87	50

Anal. 1 to 8. Eiker

Anal. 9 to 19. Grøsli

N.D. Not detected.

during metamorphism. Plimer and Finlow-Bates stated that both species could be primary in nature and that their formation was a function of sea-water depth and sulphur concentration. Shallow water (less than 500 m) and higher sulphur levels would favour pyrite, while deeper water and low sulphur levels would favour pyrrhotite. They cited the Sullivan and Broken Hill deposits, which have sedimentary host rocks, as examples of deep water deposits. On the other hand, Kuroko and Canadian Archaean deposits were considered of shallow water type, characterised by mainly volcanic host rocks. The deposits of Grøsli and Eiker have been considered to have similarities with the Kuroko-type deposits; thus the preponderance of pyrrhotite at Grøsli might be seen as a result of low sulphur species concentrations.

Analysis of the Grøsli and Eiker deposits in terms of proximal and distal characteristics is presented in Table 12.2. The table suggests very strongly that the Grøsli ores represent more distal type deposits than the Eiker ores. The strongest evidence for this conclusion is the presence of altered host rocks at Eiker while no such equivalents occur at Grøsli. The predominance of pyrrhotite at Grøsli and pyrite at Eiker also appear to be a primary feature rather than one produced by metamorphism.

12.4 CONCLUSIONS

Since the Grøsli and Eiker occurrences show many similarities with the Kuroko-type deposits, it is suggested that they were formed as a part of late stage, submarine volcanic activity of an intermediate to acidic character.

TABLE 12.2. COMPARISON OF 'PROXIMAL' AND 'DISTAL' FEATURES OF GRØSLI AND EIKER

GRØSLI			EIKER
FEATURE	RELATIVE POSITION		FEATURE
$\frac{Cu}{Cu+Zn} = 0.16$	DISTAL	PROXIMAL	$\frac{Cu}{Cu+Zn} = 0.30$
Pyrrhotite dominant	DISTAL	PROXIMAL	Pyrite dominant
No significant enrichment or depletion trends.	DISTAL	PROXIMAL	Host rocks enriched in MgO.
$\frac{Rb}{Sr} = 0.64$	DISTAL	PROXIMAL	$\frac{Rb}{Sr} = 1.51$

From Large (1977)
Plimer and Finlow-Bates (1977)
Plimer (1978)

It is suggested that the Eiker deposit formed relatively close to the exhalative centre as a type III deposit of Colley's classification. The host rocks were original volcanics, altered by hydrothermal ore solutions, which increased their magnesia content. The interbedded more-basic rocks in the Eiker area were of basaltic-andesitic nature and are considered to represent earlier differentiates of the supra-crustal sequence.

The ore body at Grøsli is considered to be a more distal type of deposit akin to the type IV deposit of Colley's classification. The predominance of pyrrhotite over pyrite suggests that this ore may have been deposited in somewhat deeper water than the Eiker deposit. The presence of probable sedimentary rocks in the sequence at Grøsli also indicates a more distal origin for this deposit.

Information about the exact relationship between these two deposits is sparse. They are both of the same age and have undergone the same metamorphic events of the Kongsberg Series block. If the two deposits are viewed as synchronous then there was an east to west succession to more acidic host rocks.

In the opinion of Jacobsen and Heier (1978), the supracrustal sequences were deposited directly as mantle derived volcanics, immediately prior to the Svecofennian orogeny, during which the rocks underwent upper amphibolite grade metamorphism. The effects of this event are preserved in the supracrustals at Eiker (the supracrustals at Grøsli did not have bulk chemistries which allowed the development of characteristic minerals). In the ore deposits, because of their readiness to recrystallize, subsequent events have obliterated the features of this metamorphic event. During the Svecofennian, amphibolite and dioritic bodies were intruded throughout the Kongsberg Series. This is therefore the earliest possible date for the intrusion of the large amphibolite bodies at Eiker.

The next major event was the intrusion of a group of gabbroic bodies. These 'Vinor' intrusives form the basic bodies at Grøsli, which intruded through (and partly assimilated) the ore deposit. At Eiker, the amphibolites may be of early 'Vinor' age. A younger metagabbro body and fine grained amphibolite dykes are of identical character to the intrusions at Grøsli. The 'Vinor' intrusive event spanned the commencement of the Mid-Amphibolite facies metamorphism of the Sveconorwegian orogeny and also followed the shearing which developed the major mylonite zone on the west of the Kongsberg Series. This shearing event also produced dislocation along the Eiker ore deposit. The intrusives at Grøsli show no cataclastic features, while those at Eiker were involved in the shearing, again possibly indicating the greater age of the Eiker intrusives.*

The 'Vinor' intrusives at Grøsli have MORB trace element characteristics, while those at Eiker have transitional IAT/MORB characters. The intrusive event is considered to be an effect of a short period of tensional plate movement prior to the Sveconorwegian orogeny. Incipient rifting allowed the intrusion of the gabbroic bodies. Commencement of the Sveconorwegian orogeny caused abortion of the rifting. There is a succession of intrusive rocks east-west across the Kongsberg Series which may be explained by the movement of a mantle plume westwards with time. The 'Vinor' intrusions occurred on the tensional side of this plume as it travelled westwards across the series. Thus, the large

---

Footnote: A N-S zone of mylonite and ultramylonite of the same relative age as that at the western margin has recently been found some 2 km east of Eiker (Starmer, Norsk.Geol.Tidssk., 1980, in press).

'Vinor' bodies could be suggested to decrease in age westwards: this is consistent with the view on the relative age of the basic rocks at Eiker and Grøsli.

At Grøsli, Sveconorwegian metamorphism produced large pyrite porphyroblasts, while self-annealed mylonites were formed in the Eiker deposit. The metamorphic event led to the development of a large grain size within both ore deposits. A succession of lower grade events then affected the Kongsberg Series. An Epidote-Amphibolite overprint was heavier in the west than in the east, with the production of epidote-rich rocks in the Grøsli area. This event also fixed the geothermometers and the geobarometers of the two ore deposits. Below this grade it is assumed that kinetics prevented any further reaction of the ores at lower temperatures and pressures.

Greenschist grade effects were of a brittle nature and shearing occurred at both Grøsli and Eiker, with fracturing of both ore and silicate. The ores at Eiker were subsequently partially recrystallized while those at Grøsli underwent no such later changes. At Eiker movement at this metamorphic grade caused rejuvenation of the higher grade Sveconorwegian shear. Fluid activity was extensive within the shear zone at this stage causing chloritization and carbonatization. At Grøsli alteration haloes were caused by minor re-solution of the ore bodies to produce chloritic rocks in the vicinity of the ore bodies, while similar re-mobilization occurred in the Eiker shear zones. Thus although the ore bodies at Grøsli and Eiker have suffered two high grade metamorphic events, late stage and lower grade deformations and alterations have largely obliterated the higher grade textures, particularly at Eiker. The present ore-silicate associations appear, at first, to be of low grade type. The successive deformations could have broken stratabound ores which were present in the original supra-crustal sequence of the Kongsberg Series.

BIBLIOGRAPHY

ANDERSON, C.A. 1969 Massive sulphides and volcanism. Econ.Geol. 64, 129-146.

ANDRAEA, M.O. 1974 Chemical and stable isotope composition of high grade metamorphic rocks from the Arendal area, southern Norway. Contrib.Mineral.Petrol. 47, 299-316.

ARNOLD, R.G. 1962 Equilibrium relations between pyrrhotite and pyrite from 325°C to 743°C. Econ.Geol. 57, 72-90.

BAER, A.J. 1976 The Grenville Province in Helikian times: a possible model of evolution. Phil.Trans.R.Soc.Lond. A280, 499-515.

BARNES, H.L. (ed.) 1967 Geochemistry of hydrothermal ore deposits. Holt, Rinehart and Winston, New York. 670pp.

BARTON, P.B. and TOULMIN, P. 1966 Phase relations involving sphalerite in the Fe-Zn-S system. Econ.Geol. 61, 815-849.

BERGE, J.W. 1978 A re-examination of the association of magnesium and massive sulphide ores. Geol.Fören.Stock. Förh. 100, 155-170.

BERTHELSEN, A. 1969 Globulith. A new type of intrusive structure exemplified by metabasic bodies in the Moss area, S.E. Norway. Norges geol.Unders. 266, 70-85.

BICKLE, M.J. and NISBET, E. 1972 The oceanic affinities of some alpine mafic rocks based on their Zr-Ti-Y contents. J.Geol. Soc. 128, 267-271.

BOORMAN, R.S. 1967 Subsolidus studies in the ZnS-ZnS-FeS_2 system. Econ.Geol. 62, 614-631.

BOORMAN, R.S., SUTHERLAND, J.K. and CHERNYSHEV, L.V. 1971 New data on the sphalerite-pyrrhotite-pyrite solvus. Econ.Geol. 66, 670-673.

BORCHERT, H. 1960 Genesis of marine sedimentary ores. Trans.Inst. Mining Met. 69, 261-279.

BUGGE, A. 1937 Flesberg og Eiker. Norges geol.Unders. 143, 118pp.

BUGGE, C. 1917 Kongsbergfeltets Geologi. Norges geol.Unders. 82, 272pp.

BUGGE, J.A.W. 1943 Geological and petrological investigations in the Kongsberg-Bamble formation. Norges geol.Unders. 160, 150pp.

BUGGE, J.A.W. 1978 In Bowie, S.H.U., Kvalheim, A. and Haslam, H.W. (eds.). Mineral deposits of Europe. Vol.1: Northwest Europe. 199-249.

CAMPBELL, F.A. and WILLIAMS, K.L. 1968 Composition of sphalerite from Quemont Mine, Quebec. Econ.Geol. 63, 824-831.

CANN, J.R. 1970 Rb, Sr, Y, Zr and Nb in some ocean-floor basaltic rocks. Earth Planet.Sci.Lett. 10, 7-11.

CARMICHAEL, I.S.E. 1964 The Petrology of Thingmuli, a Tertiary Volcano in Eastern Iceland. J.Petrol. 5, 435-460.

CARSTENS, C.W. 1919 Oversigt over Trondhjemsfeltet berbygning. Kgl.norske Vidensk.Selsk.Förh. no.1, 152pp.

CHASE, C.G. and GILMER, T.H. 1973 Precambrian plate tectonics: the mid-continent gravity high. Earth Planet.Sci.Lett. 21, 70-78.

CHERNYSHEV, L.V. and ANFILOGOV, U.N. 1968 Subsolidus phase relations in the ZnS-FeS-FeS_2 system. Econ.Geol. 63, 841-847.

CHRISTIANSEN, R.L. and LIPMAN, P.W. 1972 Cenozoic volcanism and plate tectonic evolution of the western United States. II. Late Cenozoic. Phil.Trans.R.Soc.Lond. A271, 249-284.

CLARK, B.R. and Kelly, W.C. 1973 Sulphide deformation studies: I. Experimental deformation of pyrrhotite and sphalerite to 2000 bars and 500°C. Econ.Geol. 68, 332-352.

COLLEY, H. 1976 Classification and exploration guide for Kuroko-type deposits based on occurrences in Fiji. Trans.Instn.Min. Metall. (Sect.B: Appl. earth sci.) 84, B190-B199.

COOMBS, D.S. and WILKINSON, J.F.G. 1969 Lineages and fractionation trends in undersaturated volcanic rocks from the east Otago volcanic province (New Zealand) and related rocks. J.Petrol. 10, 440-501.

COOPER, A.F. and LOVERING, J.F. 1970 Greenschist amphiboles from Haast River, New Zealand. Contrib.Mineral.Petrol. 27, 11-24.

CROSS, W. 1915 Lavas of Hawaii and their relations. U.S. geol. Surv.Prof.P. 88, 97pp.

DALY, R.A. 1911 Magmatic differentiation in Hawaii. J.Geol. 19, 289-316.

DALY, R.A. 1933 Igneous rocks and the depths of the earth. McGraw-Hill, New York.

DE BAUMONT, E. 1847 Note sur les emanations volcaniques et metal-liferes. Bull.Soc.Geol.France 4, 1249.

DEER, W.A., HOWIE, R.A. and ZUSSMAN, J. 1963 Rock forming minerals. Vol.3. Sheet silicates. Longmans, London, 270pp.

DE LONG, S.E., HODGES, F.N. and ARCULUS, R.J. 1976 Ultramafic and mafic inclusions, Kanaga Island, Alaska, and the occurrence of alkaline rocks in island arcs. Geol.Soc.Am.Bull. 87, 275-288.

DESBOROUGH, G.A. and CARPENTER, R.H. 1965 Phase relations of pyrrhotite. Econ.Geol. 60, 1431-1450.

DEWEY, J.F. and BURKE, K.C.A. 1973 Tibetan, Variscan and Precambrian basement reactivation: products of continental collision. J.Geol. 81, 683-692.

DONALDSON, J.A. and IRVING, E. 1972 Grenville front and rifting of the Canadian shield. Nature phys.sci. 237, 139-140.

EINAUDI, M.T. 1968 Sphalerite-Pyrrhotite-Pyrite equilibria - a re-evaluation. Econ.Geol. 63, 832-834.

ELLIOTT, R.B. 1973 The Chemistry of Gabbro/Amphibolite transitions in South Norway. Contrib.Mineral.Petrol. 38, 71-79.

EVANS, B.W. and LEAKE, B.E. 1960 The composition and origin of the striped amphibolites of Connemara, Ireland. J.Petrol. 1, 337-363.

FIELD, D. and ELLIOTT, R.B. 1974 The Chemistry of Gabbro/Amphibolite transitions in South Norway. II. Trace Elements. Contrib. Mineral.Petrol. 47, 63-76.

FINLOW-BATES, T., CROXFORD, N.J.W. and ALLAN, J.M. 1977 Evidence for, and implications of, a primary FeS phase in the lead-zinc bearing sediments at Mount Isa. Mineral.Deposita. 12, 143-149.

FLOYD, P.A. and WINCHESTER, J.A. 1978 Identification and discrimination of altered and metamorphosed volcanic rocks using immobile elements. Chem.Geol. 21, 291-306.

FOX, J. 1979 Host-rock geochemistry and massive volcanogenic sulphide ores. C.I.M. Bulletin 72, 127-134.

FROESE, E. and GUNTER, A.E. 1976 A note on the pyrrhotite-sulphur vapour equilibrium. Econ.Geol. 71, 1589-1594.

GARCIA, M.O. 1978 Criteria for the identification of ancient volcanic arcs. Earth Sci.Rev. 14, 147-165.

GARSON, M.S. and MITCHELL, A.H.G. 1977 Mineralization at destructive plate boundaries: a brief review. In Volcanic processes in ore genesis (London: Institution of Mining and Metallurgy and Geological Society). 81-97.

GILMOUR, P. 1971 Strata-bound massive pyritic sulphide deposits - a review. Econ.Geol. 66, 1239-1244.

GLIKSON, A.Y. 1976 Archaean to Early Proterozoic shield elements: relevance to plate tectonics. In, Strong, D.F. (ed.), Metallogeny and plate tectonics. Geol.Ass.Can.Sp.Pap. 14, 487-576.

GRAPES, R.H. and GRAHAM, C.M. 1978 The actinolite-hornblende series in metabasites and the so-called miscibility gap: a review. Lithos 11, 85-97.

GROVES, D.I., BINNS, R.A., BARRETT, F.M. and MCQUEEN, K.G. 1974 Sphalerite compositions from Western Australian nickel deposits, a guide to equilibria below 300°C. Econ.Geol. 69, 391-396.

HELOVUORI, O. 1979 Geology of the Pyhäsalmi ore deposit, Finland. Econ.Geol. 74, 1084-1101.

HENLEY, R.W. and THORNLEY, P. 1979 Some geothermal aspects of polymetallic massive sulphide formation. Econ.Geol. 74, 1600-1612.

HIETANEN, A. 1975 Generation of potassium-poor magmas in the northern Sierra Nevada and the Svecofennian of Finland. J.Res.U.S.geol. Surv. 3, 631-645.

HOFFMAN, P. 1973 Evolution of an early Proterozoic continental margin: the Coronation Geosyncline and associated aulacogens of the northwestern Canadian Shield. Phil.Trans.R.Soc.Lond. A273, 547-581.

HOBBS, B.E., MEANS, W.D. and WILLIAMS, P.F. 1976 An outline of structural geology. J. Wiley, New York. 571pp.

HUHTALA, T. 1979 The geology and zinc-copper deposits of the Pyhäsalmi-Pielavesi district, Finland. Econ.Geol. 74, 1069-1083.

HUTCHINSON, R.W. 1973 Volcanic sulphide deposits and their metallogenic significance. Econ.Geol. 68, 1223-1246.

HUTCHINSON, R.W. and HODDER, R.W. 1972 Possible tectonic and metallogenic relationships between porphyry copper and massive sulphide deposits. Canadian Mining Metall.Bull. 65, no.718, 34-40.

IRVINE, T.N. and BARAGAR, W.R.A. 1971 A guide to the chemical classification of the common volcanic rocks. Can.J.Earth Sci. 8, 523-548.

IRVING, E. and MCGLYNN, J.C. 1976 Proterozoic magneto-stratigraphy and the tectonic evolution of Laurentia. Phil.Trans.R.Soc.Lond. A280, 433-468.

JACOBSEN, S.B. 1975 Geochronological, geochemical and petrological investigations of Precambrian rocks in the Kongsberg area. Cand.real.thesis. University of Oslo.

JACOBSEN, S.B. and HEIER, K.S. 1978 Rb-Sr isotope systematics in the metamorphic rocks, Kongsberg sector, south Norway. Lithos 11, 257-276.

JENKS, W.F. 1971 Tectonic transport of massive sulphide deposits in submarine volcanic and sedimentary host rocks. Econ.Geol. 66, 1215-1224.

JOHNSTON, J.S., MUTCHLER, F.E. and BARR, D. 1976 Granites and granites - a statistical investigation of palaeostress gauges. Geol.Soc.Am., Abstr. Programs. 8, 943.

JOPLIN, G.A. 1963 Chemical analyses of Australian rocks. Part 1: Igneous and metamorphic. Dept. National Development Bureau of Mineral Resources, Geology and Geophysics. Bull. no.65.

KATADA, M. 1965 Petrography of Ryoke metamorphic rocks in northern Kiso district, central Japan. J.Jap.Assoc.Mineral.Petrol.econ. Geol. 53, 77-90, 55-164, 187-204.

KING, H.F. and THOMSON, B.P. 1953 Geology of the Broken Hill District. In: Edwards, A.B. (ed.), Geology of the Australian ore deposits. 533-577. Fifth Empire Mining and Metallurgical Congress, Australia and New Zealand.

KINKEL, A.R. 1966 Massive pyritic deposits related to volcanism and possible methods of emplacement. Econ.Geol. 61, 673-694.

KLEIN, C. 1969 Two-amphibole assemblages in the system actinolite-hornblende-glaucophane. Am.Mineral. 54, 212-237.

KNIGHT, C.L. 1957 Ore genesis - the source bed concept. Econ.Geol. 52, 808-817.

KOO, J. and MOSSMAN, D.J. 1975 Origin and metamorphism of the Flin Flon stratabound Cu-Zn sulphide deposit, Saskatchewan and Manitoba. Econ.Geol. 70, 48-62.

KULLERUD, G. 1953 The FeS-ZnS system: a geological thermometer. Norsk.Geol.Tiddsk. 32, 61-147.

KUNO, H. 1960 High-alumina basalt. J.Petrol. 1, 121-145.

LAMBERT, I.B. 1973 The features and genesis of the Kuroko-type Cu-Zn-Pb-Ag-Au deposits of Japan, with comments on some other ore deposits of volcano-sedimentary sequences. C.S.I.R.O., Div. Mineralogy Inv.Rept. 98.

LAMBERT, I.B. and SATO, T. 1974 The Kuroko and associated ore deposits of Japan: A review of their features and metallogenesis. Econ.Geol. 69, 1215-1236.

LARGE, R.R. 1977 Chemical evolution and zonation of massive sulphide deposits in volcanic terrains. Econ.Geol. 72, 549-572.

LATVALAHATI, U. 1979 Cu-Zn-Pb ores in the Aijala-Orijarvi area, Southwest Finland. Econ.Geol. 74, 1035-1059.

LEAKE, B.E. 1964 The chemical distinction between ortho and para-amphibolites. J.Petrol. 5, 238-254.

LEBLANC, M. and BILLAUD, P. 1978 A volcano-sedimentary copper deposit on a continental margin of Upper Proterozoic age: Bleida (Anti-Atlas, Morocco). Econ.Geol. 73, 1101-1111.

LE MAITRE, R.W. 1976 The chemical variability of some common igneous rocks. J.Petrol. 17, 589-637.

LUSK, J. and FORD, C.E. 1978 Experimental extension of the sphalerite geobarometer to 10 kbar. Am.Mineral. 63, 516-519.

LUSK, J., CAMPBELL, F.A. and KROUSE, H.R. 1975 Application of sphalerite geobarometry and sulphur isotope geothermometry to ores of the Quemont mine, Noranda, Quebec. Econ.Geol. 70, 1070-1083.

MACDONALD, G.A. 1960 Dissimilarity of continental and oceanic rock types. J.Petrol. 1, 172-177.

MACDONALD, J.A. 1967 Metamorphism and its effect on sulphide assemblages. Mineral Deposita. 2, 200-230.

MASON, R. 1967 Electron-probe microanalysis of coronas in a troctolite from Sulitjelma, Norway. Mineral.Mag. 36, 504-514.

MATSUKUMA, T. and HORIKOSHI, E. 1970 Kuroko deposits in Japan, a review. In: Tatsumi, T. (ed.), Volcanism and ore genesis. University Tokyo Press, Tokyo. 448pp.

MIYASHIRO, A. 1973 Metamorphism and metamorphic belts. G. Allen and Unwin. 492pp.

MIYASHIRO, A. and SHIDO, F. 1975 Tholeiitic and calc-alkaline series in relation to the behaviours of titanium, vanadium, chromium and nickel. Am.J.Sci. 275, 265-277.

NAQVI, S.M. and HUSSAIN, S.M. 1972 Petrochemistry of early Precambrian metasediments from the central part of the Chitaldrug schist belt, Mysore, India. Chem.Geol. 10, 109-135.

NOCKOLDS, S.R. 1954 Average chemical composition of some igneous rocks. Bull.geol.Soc.Am. 65, 1007-1032.

NOCKOLDS, S.R. and ALLEN, R. 1953 The geochemistry of some igneous rock series. Geochim.cosmochim.Acta. 4, 105-142.

OFTEDAHL, C. 1958 A theory of exhalative-sedimentary ores. Geol. Fören.Stockholm Förh. 80, 1-19.

OHASHI, R. 1919 On the origin of Kuroko of the Kosaka mine. J. Geol.Soc.Tokyo 26, 341-346.

O'NIONS, R.K. and HEIER, K.S. 1972 A reconnaissance Rb-Sr geochronological study of the Kongsberg area, South Norway. Norsk Geol.Tidssk. 52, 143-150.

ORVILLE, P.M. 1969 A model for metamorphic differentiation origin of thin layered amphibolites. Am.J.Sci. 267, 64-86.

PEARCE, J.A. 1979 Geochemical evidence for the genesis and eruptive setting of lavas from Tethyan ophiolites. In:

Proc.Int.Ophiolite Symposium, Cyprus (in press).

PEARCE, J.A. and CANN, J.R. 1971 Ophiolite origin investigated by discriminant analysis using Ti, Zr and Y. Earth Plant.Sci. Lett. 12, 339-349.

PEARCE, J.A. and CANN, J.R. 1973 Tectonic setting of basic volcanic rocks determined using trace element analysis. Earth Planet. Sci.Lett. 19, 290-300.

PEARCE, J.A. and GALE, G.H. 1977 Identification of ore-deposition environment from trace element geochemistry of associated igneous host rocks. In: Volcanic processes in ore genesis. (London: Institution of Mining and Metallurgy and Geological Society) 14-24.

PETRO, W.L., VOGEL, T.A. and WILBRAND, J.T. 1979 Major element chemistry of plutonic rock suites from compressional and extensional plate boundaries. Chem.Geol. 26, 217-235.

PETTIJOHN, F.J. 1957 Sedimentary rocks (2nd ed.). Harper, New York. 718pp.

PETTIJOHN, F.J. 1963 Chemical composition of sandstones - excluding carbonate and volcanic sands. U.S.Geol.Surv.Prof.P. 440-S.

PIPER, J.D.A. 1976a Definition of pre-2000 m.y. apparent polar movements. Earth Planet.Sci.Lett. 28, 470-478.

PIPER, J.D.A. 1976b Palaeomagnetic evidence for a Proterozoic supercontinent. Phil.Trans.R.Soc.Lond. A280, 469-490.

PLIMER, I.R. 1978 Proximal and distal stratabound ore deposits. Mineral.Deposita. 13, 345-353.

PLIMER, I.R. and FINLOW-BATES, T. 1978 Relationship between primary iron sulphide species, sulphur source, depth of formation and age of submarine exhalative deposits. Mineral. Deposita. 13, 399-410.

POLDERVAART, A. 1953 Metamorphism of basaltic rocks, a review. Bull.Geol.Soc.Am. 64, 259-274.

REAY A., ROOKE, J.M., WALLACE, R.C. and WHELAN, R. 1974 Lavas from Niufo'ou island, Tonga resemble ocean-floor basalts. Geology 2, 605-606.

RINGWOOD, A.E. 1970 Composition of the crust and upper mantle. Can.Tech.Rep. 1967-9, 164-165.

ROBERTS, R.G. 1975 The geological setting of the Mattagami Lake mine, Quebec: A volcanogenic massive sulphide deposit. Econ. Geol. 70, 115-129.

ROBINSON, D. and LEAKE, B.E. 1975 Sedimentary and igenous trends on AFM diagrams. Geol.Mag. 112, 305-307.

ROGERS, J.J.W. and MCKAY, S.M. 1972 Chemical evolution of geosynclinal material. In: Doe, B.R. and Smith, D.K. (eds.), Studies in mineralogy and Precambrian geology. Mem.Geol.Soc. Am. 135, 348pp. 3-28.

SANGSTER, D.F. 1972 Precambrian volcanogenic massive sulphide deposits in Canada: a review. Geol.Surv.Can.Pap. 72-22, 44pp.

SATO, T. 1974 Distribution and geological setting of the Kuroko deposits. In: Isihara et al (eds.), Geology of Kuroko ore deposits. (Tokyo: The Society of Mining Geologists of Japan) 1-9. Min.Geol.Tokyo Spec.Issue no.6, 435pp.

SATO, T. 1977 Kuroko deposits: their geology, geochemistry and origin. In: Volcanic processes in ore genesis. (London: Institution of Mining and Metallurgy and Geological Society) 153-161.

SAWKINS, F.J. 1976 Massive sulphide deposits in relation to geotectonics. Geol.Soc.Can.Spec.P. no.14, 221-240.

SAWKINS, F.J., DUNHAM, A.C. and HIRST, D.M. 1964 Iron-deficient low temperature pyrrhotite. Nature 204, 175-176.

SCOTT, S.D. 1973 Experimental calibration of the sphalerite geobarometer. Econ.Geol. 68, 466-474.

SCOTT, S.D. 1976 Application of the sphalerite geobarometer to regionally metamorphosed terrains. Am.Mineral. 61, 661-670.

SCOTT, S.D. and BARNES, H.L. 1971 Sphalerite geothermometry and geobarometry. Econ.Geol. 66, 653-669.

SCOTT, S.D. and KISSIN, S.A. 1973 Sphalerite composition in the Zn-Fe-S system below 300°C. Econ.Geol. 68, 475-479.

SMITHSON, S.B., FIKKAN, P.R. and HOUSTON, R.S. 1971 Amphibolitization of calc-silicate metasedimentary rocks. Contrib.Mineral. Petrol. 31, 228-237.

SEDERHOLM, J.J. 1916 On synantectic minerals and related phenomena. Bull.comm.Geol.de Finland.Nr. 48.

SHACKLETON, R.M. 1976 Pan-African structures. Phil.Trans.R.Soc. Lond. A280, 491-497.

SHEPPARD, S.M.F. 1977 Identification of the origin of ore-forming solutions by the use of stable isotopes. In: Volcanic processes in ore genesis. (London: Institution of Mining and Metallurgy and Geological Society) 25-41.

SIGHINOLFI, G.P. and GORGONI, C. 1978 Chemical evolution of high grade metamorphic rocks - anatexis and remotion of material from granulite terrains. Chem.Geol. 22, 157-176.

SILLITOE, R.H. 1973 Environments of formation of volcanogenic massive sulphide deposits. Econ.Geol. 68, 1321-1336.

SMIRNOV, V.I. 1968 The sources of ore-forming material. Econ. Geol. 63, 380-389.

STANTON, R.L. 1955 The genetic relation between limestone, volcanic rocks and certain ore deposits. Aust.J.Sci. 17, 173-175.

STANTON, R.L. 1972 Ore petrology. McGraw-Hill, New York, 713pp.

STANTON, R.L. 1976 Petrochemical studies of the ore environment at Broken Hill, New South Wales: 4 - environmental synthesis. Trans.Inst.Min.Metall. (Sect.B: Appl.earth sci.) 85, B221-B233.

STARMER, I.C. 1969 Basic plutonic intrusions of the Risør-Søndeled area, South Norway: the original lithologies and their metamorphism. Norsk Geol.Tidssk. 49, 403-431.

STARMER, I.C. 1977 The geology and evolution of the southwestern part of the Kongsberg series. Norsk Geol.Tidssk. 57, 1-22.

STARMER, I.C. 1979 The Kongsberg Series margin and its major bend in the Flesberg area, Numedal, Buskerud.Norges.geol.Unders. 351, 99-120.

STEPHENSON, N.C.N. 1973 The origin of amphibolite and basic granulite bands in Precambrian gneisses of the south coast of Western Australia. J.Roy.Soc.W.Aust. 56, 97-102.

SUTTON, J. 1976 Introductory remarks. In: A discussion on global tectonics in Proterozoic times. Phil.Trans.R.Soc.Lond. A280, 397-663.

TATSUMI, T. and CLARK, L.A. 1972 Chemical composition of acid volcanic rocks genetically related to formation of the Kuroko deposits. Geol.Soc.Japan.J. 78, 191-201.

TATSUMI, T. and WATANABE, T. 1971 Geological environment of the

Kuroko deposits. Soc.Mining Geologists Japan Spec.Issue 3, 216-220.

TATSUMI, T., SEKINE, Y. and KANEHIRA, K. 1970 Mineral deposits of volcanic affinity. Japan. In: Tatsumi, T. (ed.), Volcanism and ore genesis: University Tokyo Press, Tokyo, 3-47.

TATSUMI, T., TAKAGI, Y. and OTAGAKI, T. 1972 Geology of the Kuroko deposit. Joint Meeting MMIJ-AIME, Tokyo. Print no. TIB1, 1-13.

THORNTON, C.P. and TUTTLE, O.F. 1960 Chemistry of igneous rocks. I. Differentiation index. Am.J.Sci. 258, 664-684.

TOULMIN, P. and BARTON, R.B. 1964 A thermodynamic study of pyrite and pyrrhotite. Geochim.Cosmochim.Acta. 28, 641-671.

TURNER, J.S. and GUSTAFSON, L.B. 1978 The flow of hot saline solutions from vents in the sea floor - some implications for exhalative massive sulphide and other ore deposits. Econ.Geol. 73, 1082-1100.

VAN DE KAMP, P.C. 1969 Origin of amphibolites in the Beartooth Mountains, Wyoming and Montana: New data and interpretation. Bull.Geol.Soc.Am. 80, 1127-1136.

VAN DE KAMP, P.C., LEAKE, B.E. and SENIOR, A. 1976 The petrography and geochemistry of some Californian arkoses with application to identifying gneisses of metasedimentary origin. J.Geol. 84, 195-212.

VAN SCHMUS, W.R. 1976 Early and Middle Proterozoic history of the Great Lakes area, North America. Phil.Trans.R.Soc.Lond. A280, 605-628.

VOKES, F.M. 1966 On the possible modes of origin of the Caledonian sulphide ore deposit at Bleikvassli, Norland, Norway. Econ.Geol. 61, 1130-1139.

VOKES, F.M. 1969 A review of the metamorphism of sulphide deposits. Earth Sci.Rev. 5, 99-143.

VOKES, F.M. and GALE, G.H. 1976 Metallogeny relatable to global tectonics in Southern Scandinavia. Geol.Assoc.Can.Spec.P. no.14, 414-441.

WEDEPOHL, K.H. 1969 Composition and abundance of common igneous rocks. In: Wedepohl, K.H. (ed.), Handbook of geochemistry. I. Springer, Berlin, 442pp, 227-249.

WILCOX, R.E. and POLDERVAART, A. 1958 Metadolerite dyke swarm in the Bakerville-Roan Mountain area, North Carolina. Bull.Geol. Soc.Am. 69, 1323-1368.

WILKINSON, J.M. and CANN, J.R. 1974 Trace elements and tectonic relationships of basaltic rocks in the Ballantrae igneous complex, Ayrshire. Geol.Mag. 11, 35-41.

WILLIAMS, D. 1960 Genesis of sulphide ores. Proc.Geol.Assoc. 71, 245-311.

WINDLEY, B.F. 1973 Crustal development in the Precambrian. Phil. Trans.R.Soc.Lond. A273, 321-341.

WINDLEY, B.F. 1977 The evolving continents. J. Wiley, New York, 385pp.

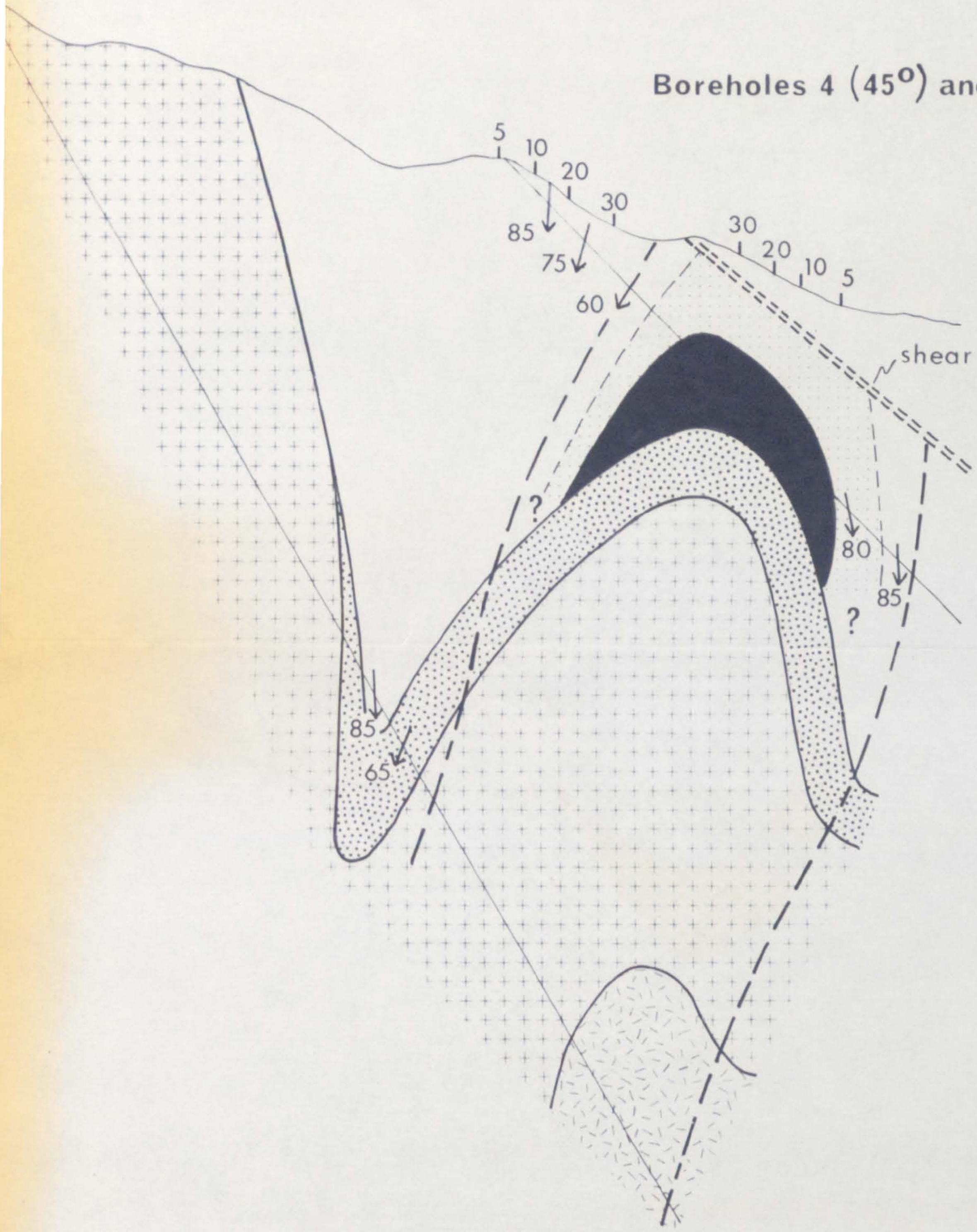
WINKLER, H.G.F. 1976 Petrogenesis of metamorphic rocks (4th ed.). Springer-Verlag, New York, 334pp.

WYNNE-EDWARDS 1976 Proterozoic ensialic orogenesis: the millepede model of ductile plate tectonics. Am.J.Sci. 276, 927-953.

PULLOUTS

FIG. 2.3.5

Boreholes 4 (45°) and 5 (60°)



Borehole 8 (45°)

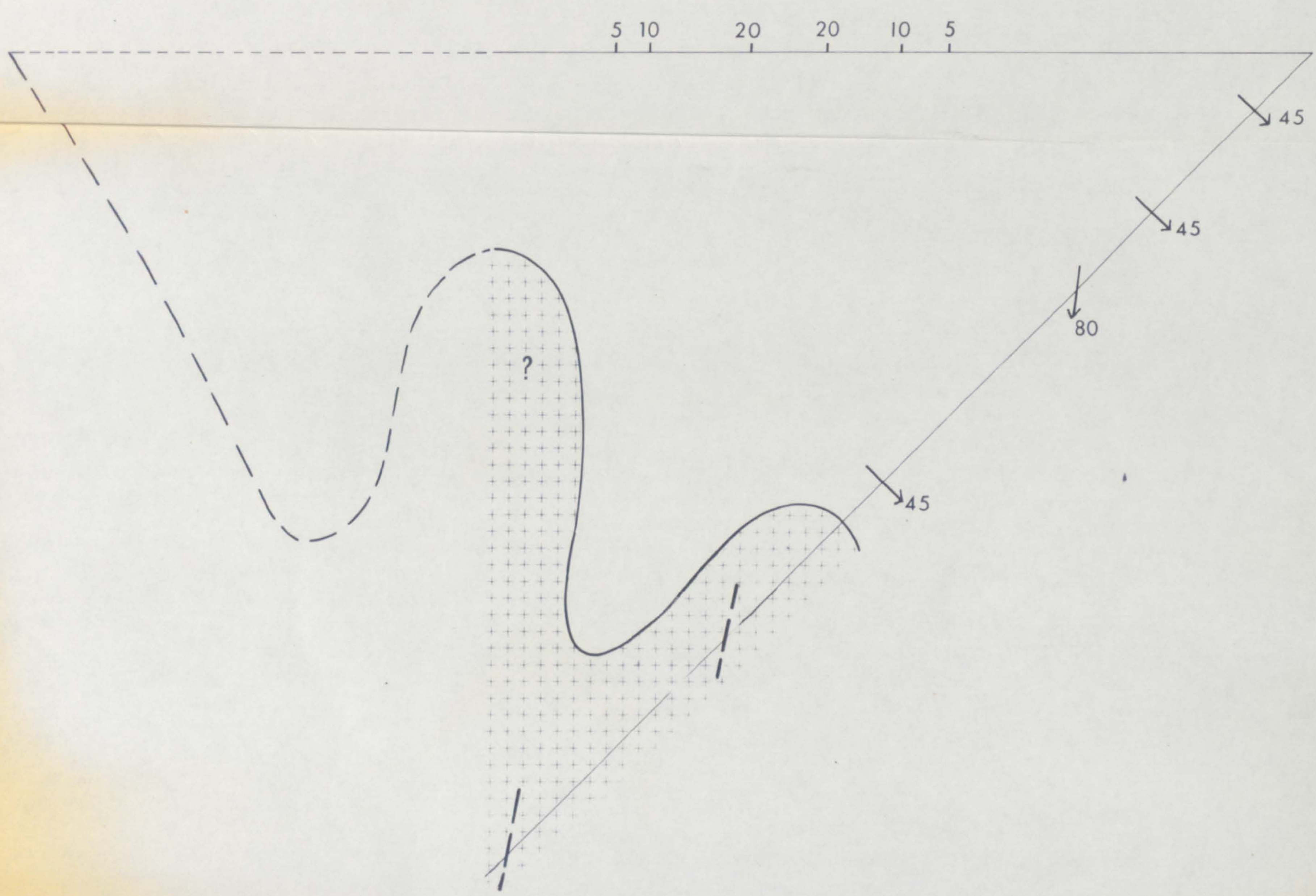


FIG. 2.3.1

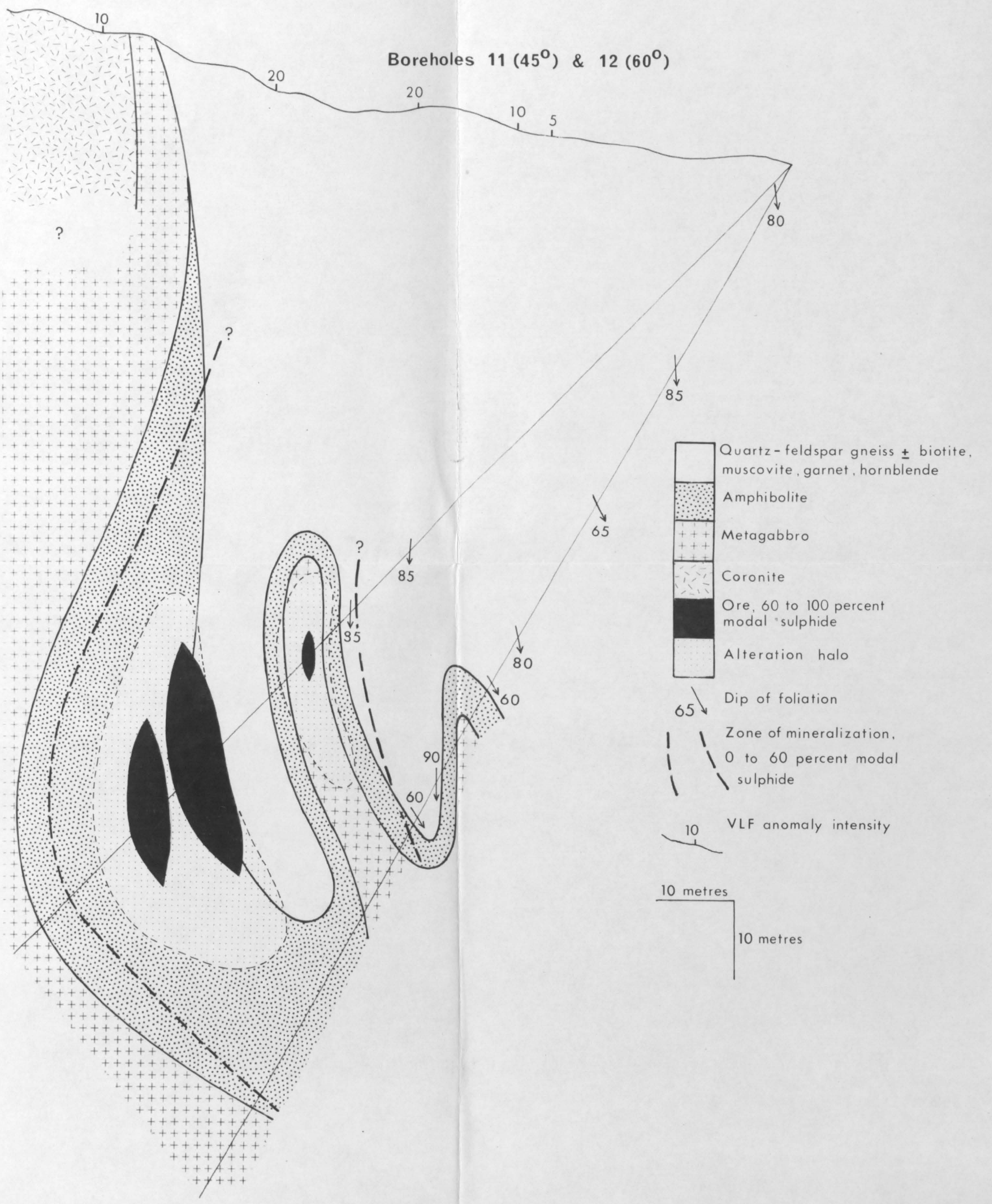
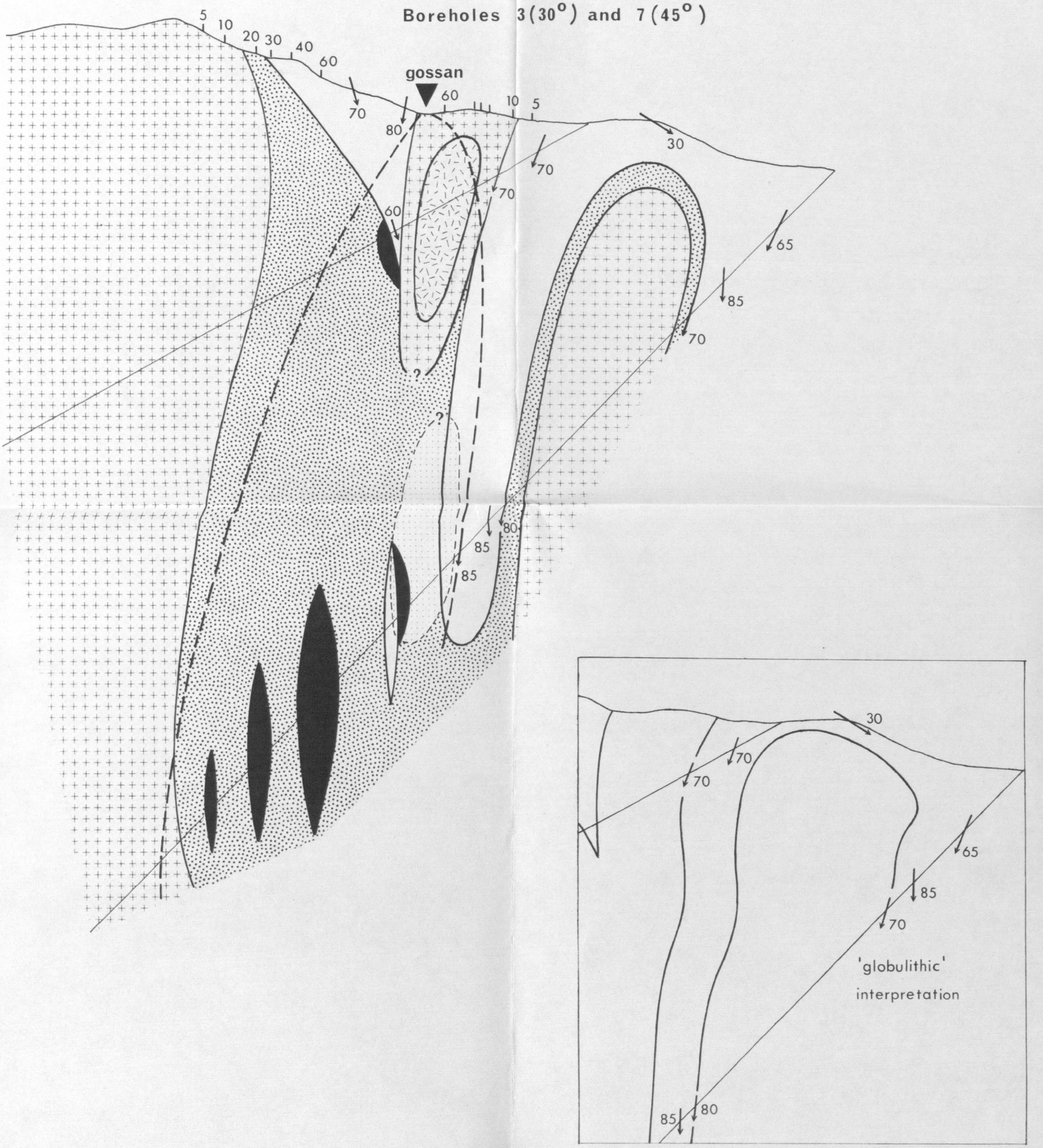
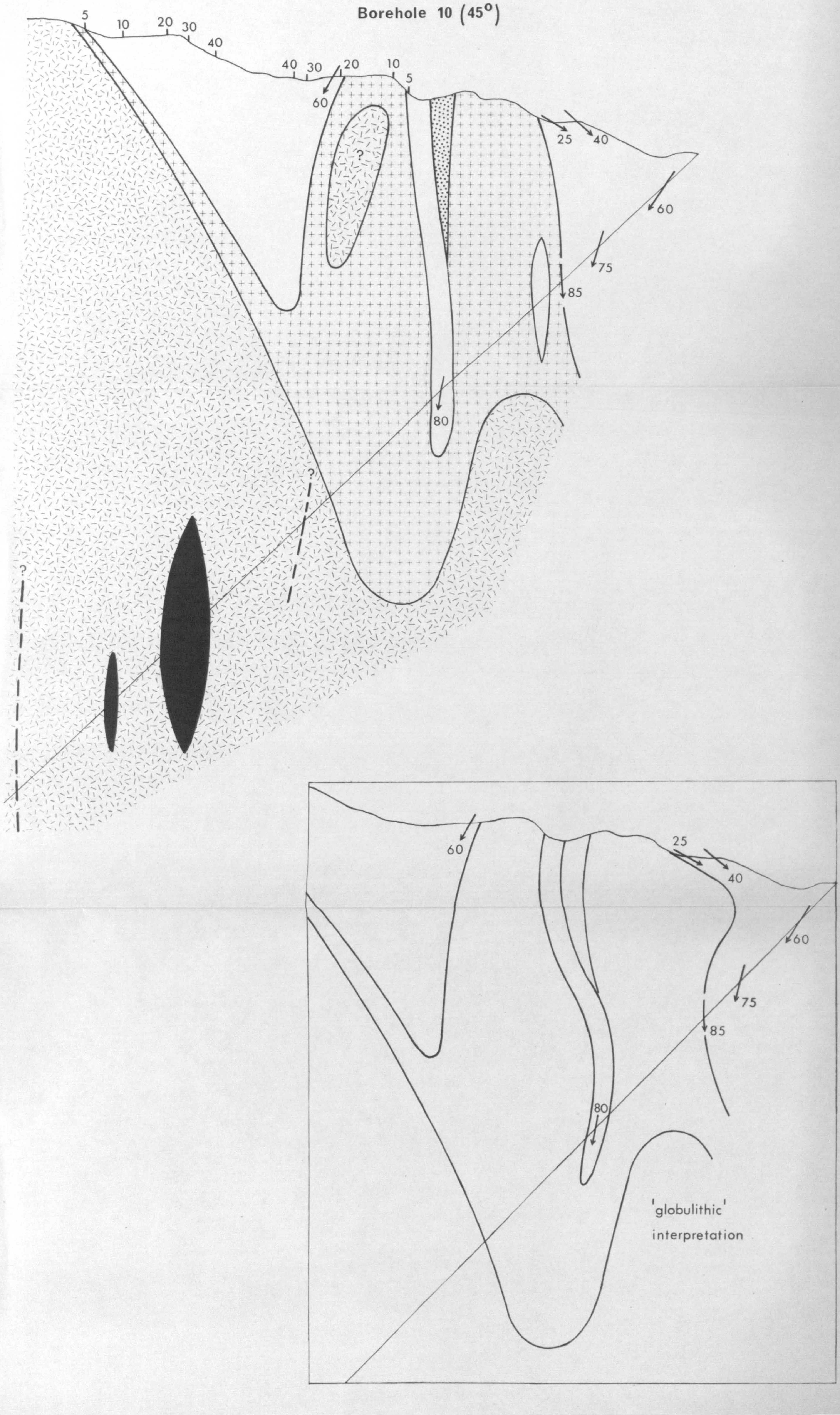

Boreholes 11 (45°) & 12 (60°)
10
20
20
10
5
80
85
65
85
85
80
60
90
60
?
?
?
Quartz - feldspar gneiss ± biotite, muscovite, garnet, hornblende
Amphibolite
Metagabbro
Coronite
Ore, 60 to 100 percent modal sulphide
Alteration halo
Dip of foliation
65
Zone of mineralization, 0 to 60 percent modal sulphide
VLF anomaly intensity
10
10 metres
10 metres

FIG. 2.3.2



Boreholes 3(30°) and 7(45°)
gossan
5
10
20
30
40
60
70
80
60
10
5
30
70
70
60
65
85
70
?
?
85
80
85
30
70
70
65
85
70
85
80
'globulithic'
interpretation

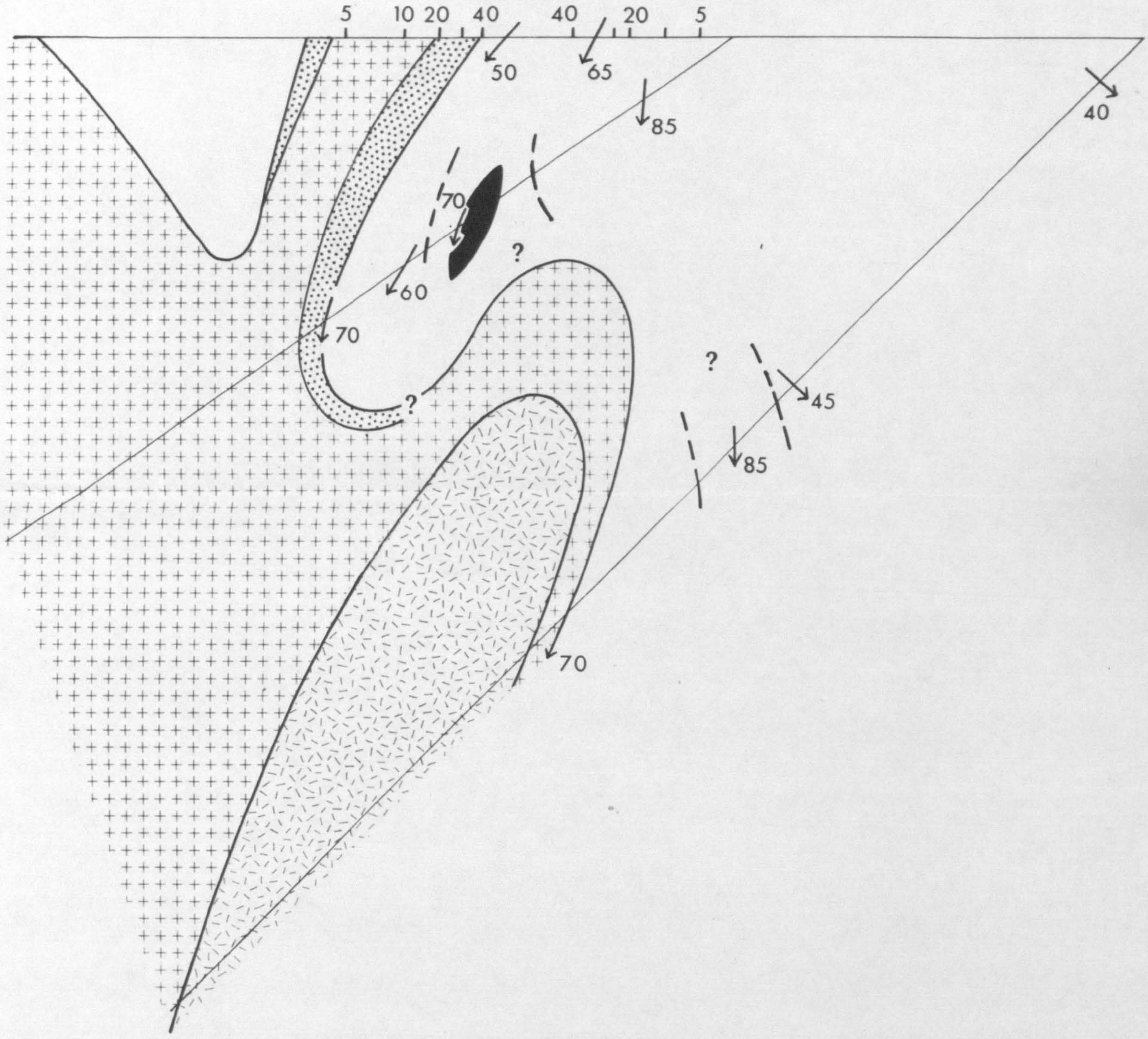
FIG. 2.3.3



Borehole 10 (45°)
5
10
20
30
40
40
30
20
10
5
60
?
25
40
60
75
85
80
?
?
60
25
40
60
75
85
80
'globulithic'
interpretation

FIG. 2.3.4

Boreholes 2 (30°) and 6 (45°)



Borehole 9 (45°)

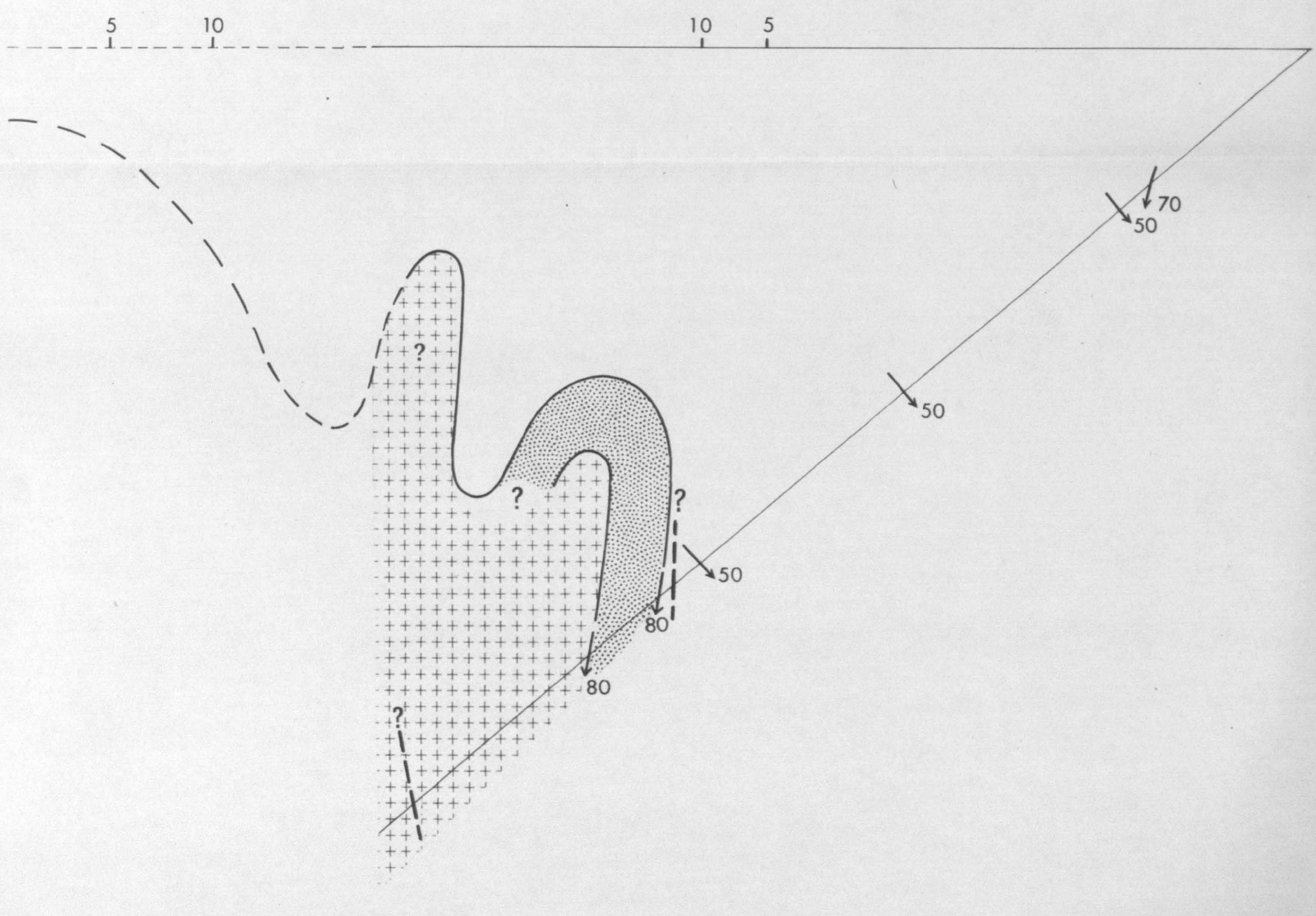


FIG. 2.3.6

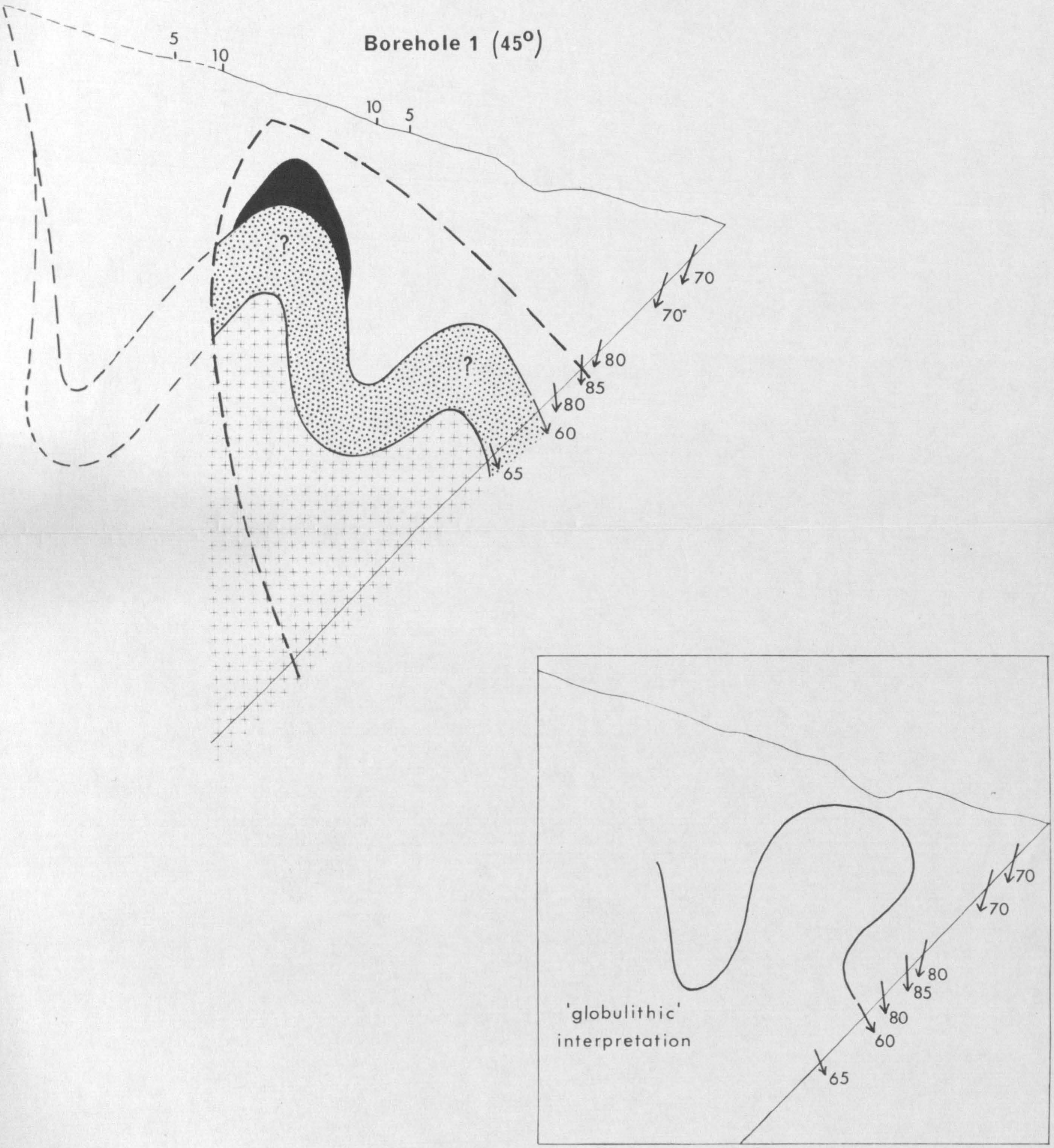

Borehole 1 (45°)
5
10
10
5
?
?
70
70
80
85
80
60
65
70
70
80
85
80
60
65
'globulithic'
interpretation

FIG. 2.4

Isometric block diagram showing relative positions of gneiss-'Vinor' junctions in borehole sections.

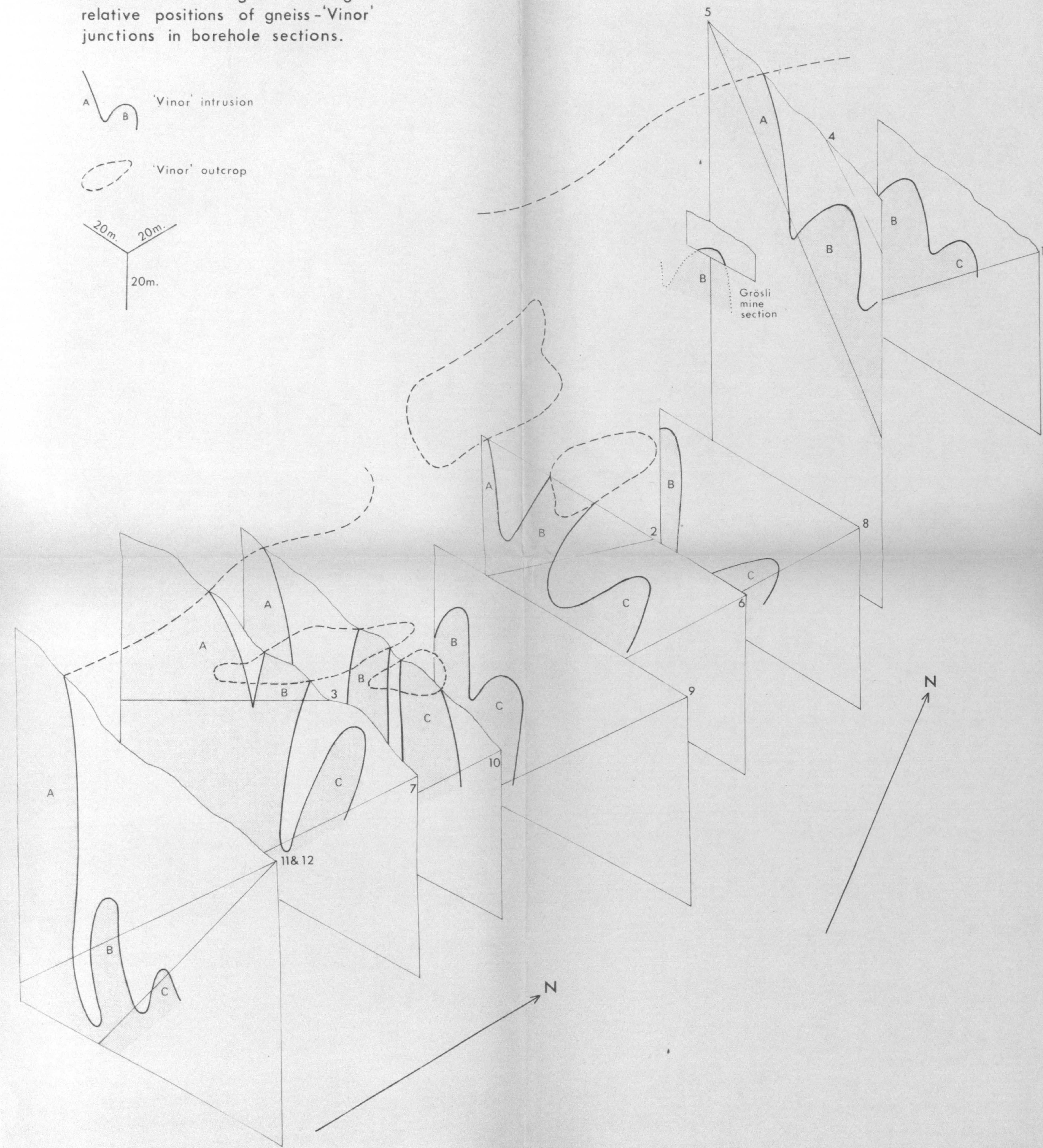


FIG. 2.2 VLF anomaly survey

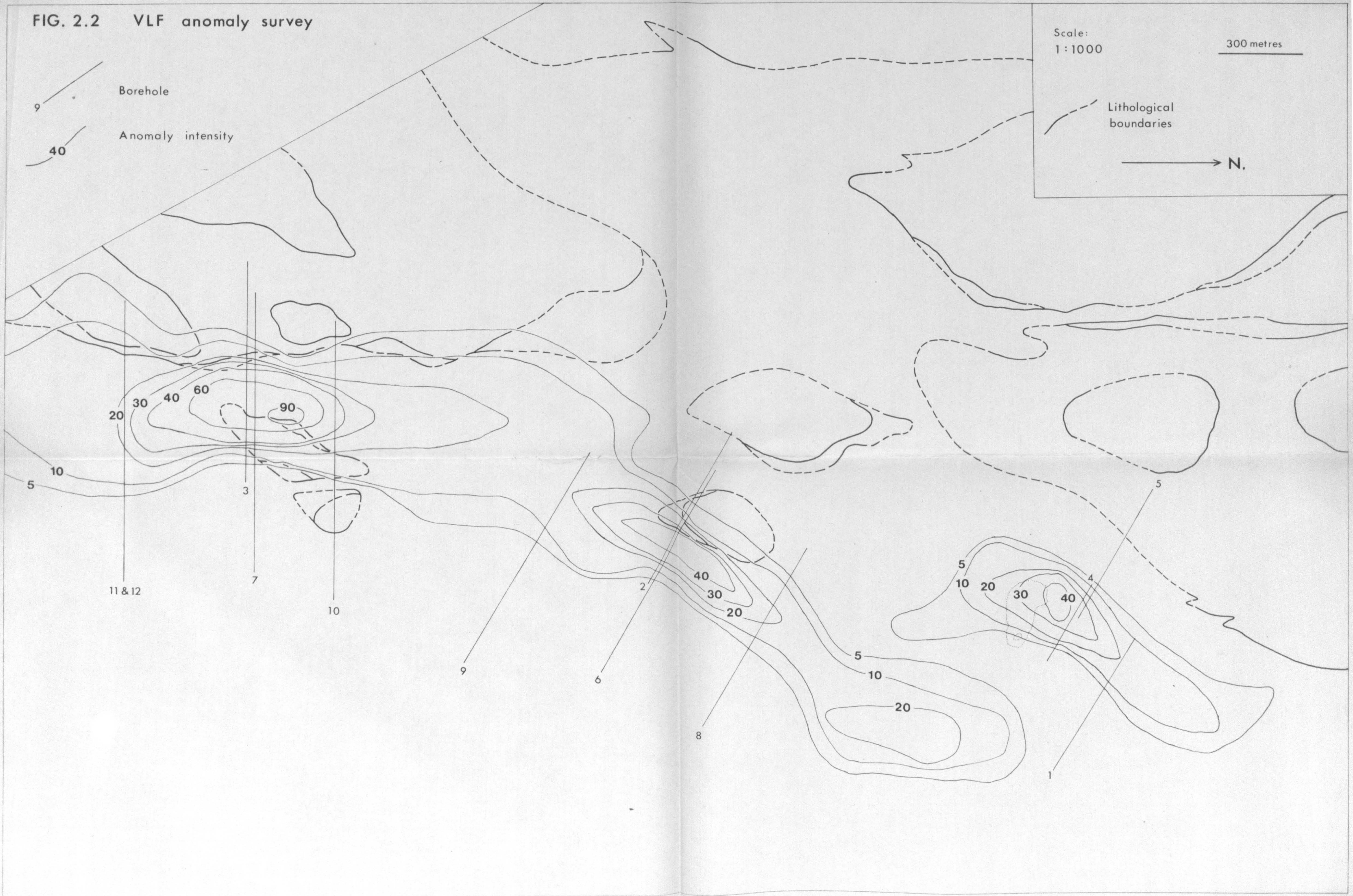


FIG. 6·2 STRUCTURE OF THE EIKER MINE AREA

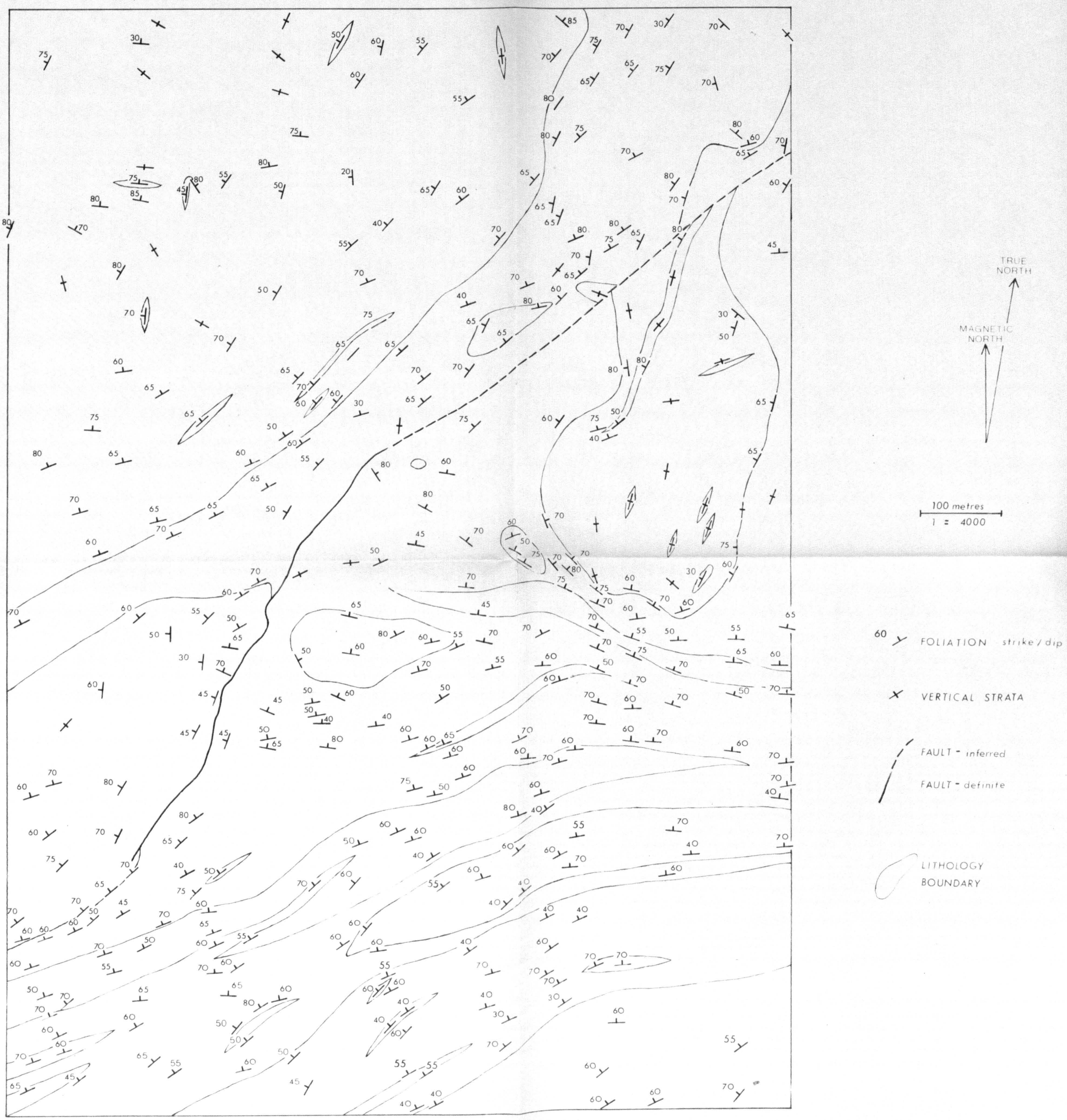


FIG. 6·1

LITHOLOGIES OF THE EIKER MINE AREA.

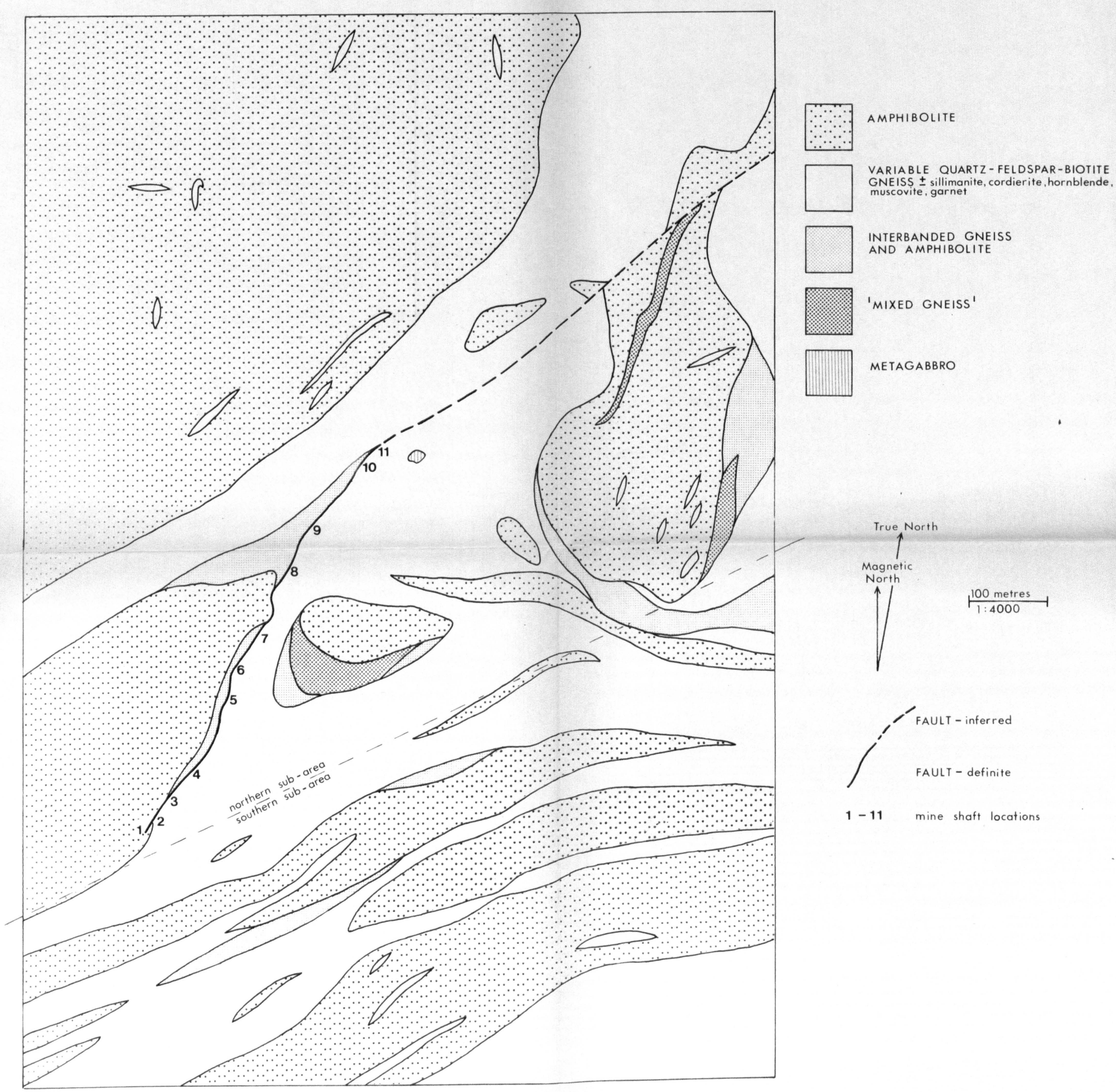


FIG. 6·3·1

Shaft 8
Shaft 9
Shaft 10
Shaft 11
Upper shear
Lower shear
40
50
70
60
45
35
80
65
55
75
15

FIG. 6·3·3


Upper shear
Lower shear
Shaft 2
Shaft 3
Shaft 4
working 1
SHEARS BECOME DIFFUSE TO NORTH - EAST
SHEARS DIFFUSE INTO BELT OF FRACTURED ROCK TO SOUTH - WEST

FIG. 6·3·2

GEOLOGY OF THE EIKER MINE SHEAR ZONE.

FIG. 2.1

GRÖSLI MINE AREA.

Isometric block diagram and map of surface geology.